Carbonate-siliciclastic interactions: tertiary examples from Spain

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Carbonate-siliciclastic interactions: Tertiary examples from Spain

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No part of this thesis has been previously submitted for a degree in this or any other university. The work described in this thesis is entirely that of the author, except where reference is made to previous published or unpublished work.

Abstract

It is a common misconception that shallow water biogenic carbonate development is inhibited in areas of active siliciclastic input. However, an increasing number of examples of ancient and modern coral communities are being identified which are affected by siliciclastic input and are developing in areas traditionally regarded as unfavorable. Corals can develop in nearshore settings affected by high sedimentation rates, turbidity, mobile substrates and episodic (storm-related) discharges of freshwater and terrestrially-derived sediments. Coral reefs in these systems are not necessarily impacted reefs, and represent natural states of development with coral abundance and diversity comparable to clear-water systems. Accepted models of shallow water carbonate production in clear water conditions only represent one end-member in a diverse range of shallow coral-dominated communities.

In order to investigate the development of shallow water, biogenic carbonate development under the influence of siliciclastic input, two mixed carbonate-siliciclastic successions have been studied from the Tertiary of Spain. The Vic Basin (NE Spain) and the Fortuna Basin (SE Spain) provide contrasting examples of coral reef development within siliciclastic shallow marine shelf environments bordering temperate-humid and semi-arid land areas respectively. The methods employed in this investigation were 1) high resolution sedimentary logging and sample collection, 2) petrographic and palaeontological analysis of samples and 3) quantification of non-carbonate content through acid digestion. The correlation of logged sections, and development of a facies scheme for each study area, has enabled the temporal and spatial relationships between carbonate development and siliciclastic sedimentation to be deciphered.

The Calders study area is situated within the Vic Basin. The broad environment of deposition was a moderate energy, northward-prograding siliciclastic shelf where high sediment input and unstable substrates inhibited the development of sessile calcareous biota. Carbonate development occurred following abandonment of the siliciclastic substrate. Abandonment facies, which developed on dune foresets and topsets, are dominated by large benthic foraminifera and coralline algae. Carbonate development, as high-energy foralgal shoals and muddy coral-dominated sediments, occurred down-slope. Carbonate units developed as very low angle clinoforms. Coral development was variable, existing as metre-scale patch reefs with associated debris and protected low-energy environments. Coral framework was only locally evident. Six isolated carbonate intervals are identified within the siliciclastic-dominated succession. The change from carbonate development to siliciclastic deposition is abrupt.

It is proposed that sediments situated stratigraphically above the main Calders section in the Sant Amanc area are part of the Terminal Complex. The broad environment of deposition was a protected, inner-shelf that developed in the latest stages of marine sedimentation in the Pyrenean foreland prior to regional deposition of continental sediments. Nutrient-rich conditions persisted due to terrigenous input and terrestrial run-off. Metre-scale patch reefs developed through the baffling effect of the perennial seafloor vegetation that acted as a substrate for large benthic foraminifera.

The Altorreal study area is situated within the Fortuna Basin. The environment of deposition was a marginal marine, high-energy fan delta system. Only robust organisms such as oysters were able to tolerate the periodic high-magnitude input of

coarse-grained siliciclastics. The temporary abandonment of fan-delta lobes provided a site for coral colonization. Abandonment facies are siliciclastic packstones with laminar stromatolites in up-slope areas. Carbonates contain minor siliciclastic material and form laterally restricted sigmoidal units. Corals formed a framework, with variations in coral morphology a function of water depth. Carbonate production was halted through the combined affects of freshwater input, emergence and erosion. Eroded carbonate bodies were buried by fan-delta sediments as the locus of sedimentation changed.

The main impacts of sediment input on photoautotrophs are physical burial, reduction in rates of photosynthesis though increased turbidity and changes in seawater chemistry, particularly salinity and nutrients. The amount and grainsize of siliciclastic sediment input has influenced the biota in both study areas. Siliciclastic-dominated sediments contain a low diversity fauna dominated by echinoids, molluscs and burrowing organisms. The reduction of siliciclastic input created a new environment that was initially colonized by organisms acting as r-strategists. These transitional settings are dominated by larger benthic foraminifera at Calders. At Altorreal, a prolonged period of non-deposition led to the formation of a hardground. Coral development at Calders occurred during a constant input of clay to silt-grade siliciclastics. In low-energy areas where sediment input was particularly high, constratal growth of branching corals is inferred. In marginal marine reef areas at Altorreal, coarse lithoclastic grains supported stick-like coral branches. Demise of coral communities is attributed to a number of factors. At Altorreal, emergence and erosion of the reef is inferred from erosional contacts. The development of columnar stromatolites is associated with a prolonged period of non-deposition and possibly a change in seawater chemistry that was detrimental to corals.

The development of coral communities in siliciclastic settings can be aided through the existence of a protection mechanism. At Calders and Altorreal, temporarily abandoned siliciclastic substrates provided sites away from silts of high siliciclastic input. Autogenic factors such as delta lobe switching were the most important controls on coral development in the studied areas although allogenic factors such as climate and the tectonic regime influenced rates, magnitude and grainsize of input were also important.

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"The heavier the burden, the closer our lives come to the earth, and the more real and truthful they become."

Milan Kundera, The Unbearable Lightness of Being

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1. Introduction

1.1 Aims of study

Deposits of Eocene age from NE Spain and of Miocene age from SE Spain allow the study of carbonate systems strongly influenced by siliciclastic input. The presence of these carbonate systems within siliciclastic-dominated environments is widely recognised. However, few studies evaluate the effects of siliciclastic sediment input on facies development, biota and evolution of the environment.

The primary objectives of this research are:

1. To determine the depositional environments of the deposits from the Calders area (Vic Basin, NE Spain) and the Altorreal area (Fortuna Basin, SE Spain), and to construct a detailed facies model for their evolution.
2. To assess the effects of siliciclastic input as a control on the nature and development of carbonate producing organisms within siliciclastic-dominated environments.
3. To compare the characteristics of mixed carbonate-siliciclastic systems within subtropical humid and semi-arid settings.
4. To compare the mixed carbonate-siliciclastic settings of the study areas with similar modern and ancient settings.

To meet these aims this study combines detailed field logging and sampling of carbonate and siliciclastic lithologies within both study areas with detailed petrographic, palaeontological and acid digestion analysis. Regional literature studies were undertaken to establish the broader context of the carbonate systems studied.

1.2 Outline of thesis

Chapter 1: Brief introduction to the thesis, including the aims and objectives of the study.

Chapter 2: Geological context of the Calders area in NE Spain and the Pyrenees. The tectonic and sedimentary evolution of the SE Pyrenean Foreland Basin is discussed, with particular emphasis on the Tertiary. The chronology and lithostratigraphic terms for the Eocene of the Pyrenean Foreland are described.

Chapter 3: Detailed description and environmental interpretations of facies from the Calders and Sant Amanc study area. A facies scheme is proposed for Eocene (Bartonian) inner-shelf mixed carbonate siliciclastic environments. The temporal and spatial evolution of facies is described and evaluated.

Chapter 4: Geological context of the Altorreal study area in SE Spain and the Betic Cordillera. The tectonic and palaeogeographic evolution of the western Mediterranean region from the Cretaceous to Recent is discussed, with particular emphasis on the Miocene.

Chapter 5: Detailed description and environmental interpretation of facies of the Altorreal study area, Fortuna Basin. A detailed facies scheme is proposed for Tortonian/Messinian inner shelf mixed carbonate siliciclastic environments. The temporal and spatial evolution of facies is described.

Chapter 6: An assessment of the effects of siliciclastic sediment input on biota in the Calders study area.

Chapter 7: A comparison of the studied sediments from the Calders and Altorreal areas with comparable modern and ancient examples. This chapter presents a detailed discussion of the characteristics of carbonate production within siliciclastic-dominated environments. The features of carbonate systems within different tectonic, climatic and hydrodynamic settings will be discussed.

Chapter 8: Summary and future work

Appendices:

Appendix 1: A review of the techniques and terminology used during this study.

Appendix 2: A review of benthic foraminifera palaecology, focusing on diagnostic genera present in the study areas.

Appendix 3: Tabulated thin data data, Calders and Sant Amanc study area

Appendix 4: Tabulated thin section data, Altorreal study area

Appendix 5: Sedimentary logs, Calders and Sant Amanc study area

Appendix 6: Sedimentary logs, Altorreal study area

2. Regional Geology and Field Areas, Northern Spain

2.1 Introduction

The aim of this chapter is to summarise previous work on the regional geological and palaeoenvironmental evolution of the Pyrenees, in particular the southeastern areas. In order to put the studied sections in the Vic Basin into regional context, this chapter provides a detailed review of the South Pyrenean Foreland and Vic Basin stratigraphy.

2.2 Regional Geology of the Pyrenean Region

The geology of the Pyrenean region has been influenced by a series of major structural events (**Figure 2.1**). These are the Hercynian orogeny, which produced major NE-SW structural lineaments reactivated during later tectonic episodes (Gisbert *et al.* 1983) and the break-up of Pangea (and the formation of the Tethys Ocean). The present tectonic configuration is attributed to oblique collision and shearing of the Iberian microplate with the European plate during the Late Cretaceous to Miocene Alpine Orogeny (Muñoz 1992, Puigdefàbregas and Souquet 1986, Masson and Miles 1984). Although collision is related to the Alpine Orogeny, there is no continuity between the Alpine Mediterranean and Pyrenean mountain belts, and there are distinct differences in deformation style between the Pyrenees and other parts of the Alpine orogenic belt (Banda and Wickham 1986). Therefore, the collisional event in the Pyrenees is referred to herein as the Pyrenean Orogeny.

Between the Hercynian and Pyrenean Orogenies, sedimentation in the Pyrenees was near continuous. The Pyrenees is an example of a polyhistory basin where successive stages of development are recorded as Mesozoic and Tertiary tectono-sedimentary cycles (Puigdefàbregas and Souquet 1986, Kingston *et al.* 1983). The chronology of these cycles is summarized in **Figure 2.1**.

The pre-Hercynian tectonic setting of the Pyrenees is uncertain although Zwart (1986, 1979) provides an exhaustive review of Hercynian geology. The first grouping of tectono-sedimentary cycles (Cycles 1 to 4, **Figure 2.1**) corresponds to major periods of intra-continental rifting, lasting from the Permian to the Early Cretaceous. Extension was controlled by inherited E-W, NE-SW and NW-SE orientated fractures, and led to the formation of assymmetric fault-bounded basins that existed throughout the Permian (Gisbert *et al.* 1983). Rocks of Permian age include alluvial fan

Pyrenean Cycles	Age	Basin Types	Tectonics
10	Miocene-Oligocene *~34-5 Ma*	Last stage foreland basins	
9	Eocene *~55-34 Ma*	Turbidite to fluvial fill of the migrating foreland	Thrust sheet emplacement
8	Palaeocene *~65-55 Ma*	Transition to foreland basins	
7	Late Santonian-Maastrichtian *~85-65 Ma*	Wrench basin including local folding and uplift with submarine and subaerial erosion	Initial collision in the eastern Pyrenees
6	Middle Cenomanian-Middle Santonian *~93-35 Ma*	Deeper turbidite wrench basin and related back-stepping carbonate platforms.	Wrenching
5	Middle Albian-Early Cenomanian *~110-94 Ma*	Strike-slip turbidite troughs along the North Pyrenean Fault Zone. Initial turbidite deposition.	Sinistral wrenching
4	Early Aptian-Albian *~115-110 Ma*	Rhombic sub-basins in a NW-SE trending rift system along inherited basement directions (Parentis, Adur and Pyrenees)	Rifting of the Bay of Biscay
3	Neocomian-Marremian *~140-115 Ma*	Unstable platform	
2	Lias-Malm ~210-240Ma Triassic *~250-210 Ma*	Synrift alluvial deposits to carbonate platform controlled by normal faulting along NE-SW inherited basement	Ligurian Tethys and Atlantic rifting
1	Permian *>250Ma*	Interior fracture basins	Intra-continental rifting

Convergence

Transpression

Continental Break-up

E-W progressive emergence

Rotation of Iberia

Figure 2.1 Summary of Pyrenean Permian to Miocene Pyrenean tectono-sedimentary cycles. Cycles 1 to 4 (shaded) represent a phase of intracontinental rifting. The final rifting phase was marked by accumulation of Cycle 5. Cycles 6 and 7 (shaded) were deposited during the change from a strike-slip to a convergent tectonic regime. Cycles 8 to 10 formed under a convergent tectonic regime. After Puidefàbregas and Souquet (1986). See text for discussion.

conglomerates, red mudrocks, local slope breccias and volcanoclastics, with associated andesites, rhyolites and basalts (Puigdefàbregas and Souquet 1986 and references within).

Extension continued through the Triassic and Jurassic. Triassic sediments are red conglomerates and mudrocks (interpreted as braided fluvial systems) that grade upward into fine-grained lagoonal and evaporite deposits (Puigdefàbregas and Souquet 1986 and references within). Puigdefàbregas and Souquet (1986) interpret this as a classic intra-continental rift succession. By the early Jurassic, a carbonate platform influenced by eustatic fluctuations extended over most of the Pyrenean region (Peybernes 1976, Faure 1980, 1984). By the Middle Jurassic (Dogger), shallow water carbonate development was restricted by differential subsidence related to faulting to a major horst structure known as the Occitan High (Puigdefàbregas and Souquet 1986). Shallow marine deposits covered the Occitan High while deeper hemipelagic carbonates accumulated in the adjacent basins of the eastern and western Tethyan domains (Peybernes 1976, Faure, 1980, 1984; Puigdefàbregas and Souquet 1986). A deeper water area had formed to the west by the Late Jurassic (Malm), indicated by condensed ammonite intervals and hemipelagic shales and mudrocks exposed in the western Pyrenees. Eastern areas of equivalent age are characterised by shallow marine carbonates (including limestone-dolomite complex collapse breccias and dark lagoonal limestones and dolomites) exposed in the Central and Eastern Pyrenees (Peybernes 1976, Faure, 1980, 1984; Puigdefàbregas and Souquet 1986).

The change from extensional to transtentional tectonics at the base of the Cretaceous is associated with a sudden relative sea level fall, resulting in discontinuous sedimentation and localised erosion surfaces (Puigdefàbregas and Souquet 1986). Fracturing and erosion of the Jurassic carbonate platform was followed by the deposition of a polymict carbonate breccia exposed in the central-eastern Pyrenees. A NW-SE trending rift system evolved in response to initial rifting in the Bay of Biscay and the neighbouring Cantabrian and Iberian mountain chains, and created the Parentis, Adour and Pyrenees sub-basins. These basins increased in areal extent in the Early Cretaceous (Neocomian) as a response to relative sea level rise, although sedimentation rates were still low and influenced by the inherited fracture systems (Puigdefabregas and Souquet 1986). In the Tethyan domain to the east, relative sea level rise is inferred from the development of a westward-onlapping eustatically influenced carbonate platform (the Urgonian Limestones). A Tethys-

Atlantic Ocean connection via the rift system during the Early Cretaceous is inferred from the lateral extent of marine facies (Peybernes 1982, Puigdefàbregas and Souquet 1986).

Middle Albian-early Cenomanian sea-floor spreading in the North Atlantic induced rotation and a change in a trajectory of Iberia, leading to the final break-up of the Iberian and European plates (Puigdefabregas and Souquet 1986). Tectono-sedimentary Cycle 5 (**Figure 2.1**) was deposited during the tectonic activity that caused crustal stretching along the North Pyrenean Fault Zone (NPFZ), together with the development of a deep wrench basin system (Debroas 1987). Crustal stretching at the NPFZ was accompanied by alkaline magmatism, thermal metamorphism and diapirism of Triassic evaporates (Puigdefàbregas and Souquet 1986 and references within). Progressive basin extension in the Mid- to Late-Cretaceous is associated with the deposition of retrogradational turbidites in subsiding troughs with minor carbonate platform development on the basin margins (Souquet 1985). Puigdefàbregas *et al.* (1986) infer a diachronous change from transtentional to transpressional tectonics.

The next grouping of tectono-sedimentary cycles (Cycles 6 and 7) was deposited during the gradual change from strike-slip displacement to oblique plate convergence during the Mid- to Late-Cretaceous (**Figure 2.1**). The change in tectonic regime is interpreted as a direct result of the oblique collision of the Iberian and Eurpoean plates during the Pyrenean Orogeny. Continent-continent collision resulted in the northward subduction of the lower crust and lithospheric mantle of the Iberian plate below the more stable European plate (ECORS-Pyrenees Team 1988, Roure *et al.* 1989, Muñoz 1992). The onset of deformation was slightly earlier in the east than the west, and initial collision resulted in emergence of compressive ridges in the eastern Pyrenees (Plaziat 1981, Puigdefàbregas and Souquet 1986). The Late Cretaceous global sea level fall led to the exposure and erosion of the ridges and widespread deposition of alluvial to shallow marine successions known as known as the Garumnian Facies (Puigdefàbregas *et al.* 1986, Puigdefàbregas *et al.* 1992). Fluvial and lacustrine red beds of the Garumnian Facies occur in the central and eastern Pyrenees, both to the north and south of the Axial Zone of the Pyrenean Orogen (Puigdefàbregas *et al.* 1992). The Garumnian beds overlie Hercynian basement in the north and Triassic sediments in the south (Jurado 1988). Deep-water carbonate, siliciclastic and turbidite sedimentation similar to the earlier Mesozoic wrench basins prevailed in the west until the late Paleocene (Pujalte *et al.* 1989).

Convergent plate conditions also characterise the fourth grouping of tectono-sedimentary cycles (Cycles 8, 9 and 10) (**Figure 2.1**). The transition from deep marine wrench basin to foreland basin occurred during the Paleocene (Puigdefàbregas and Souquet 1986, Puigdefàbregas *et al.* 1986). Migrating foreland basins were fully established by the early Eocene, and their development was directly related to thrust sheet emplacement (Puigdefàbregas *et al.* 1986, Burbank *et al.* 1992). The preceding wrench basin was systematically incorporated into the growing thrust-belt of the mountain chain. The east-west trending orogen consists of a double-verging wedge that developed by propagating thin-skinned, linked thrust systems in foreland areas (Puigdefàbregas *et al.* 1992 and references within). Progressive unconformities developed during the Late Oligocene-Miocene in response to the final sub-aerial emergence of the Pyrenean thrust sheets. A siliciclastic wedge comprising alluvial fan, fluvial and lacustrine sediments accumulated ahead of the final thrust sheet (Puigdefàbregas and Souquet 1986, Turner *et al.* 1984). The tectono-sedimentary evolution of the foreland basin is described in more detail in **Section 2.3.2**.

2.3 The Pyrenean Foreland Basin System

From north to south, the Pyrenean Orogen comprises the Aquitain Retro-Foreland Basin, the Northern Thrust Wedge, the Axial Zone, the South Pyrenean Thrust System and the South Pyrenean Foreland Basin (**Figure 2.2**) (Puigdefàbregas and Souquet 1986, Vergés *et al.* 1995). The southeastern and southwestern margins of the foreland are occupied by the Catalan Coastal Range and Iberian Range respectively. The Vic Basin study area is situated in the undeformed southeastern part of the Pyrenean Foreland Basin.

2.3.1 The North Pyrenean Thrust System and the Aquitain Retro-Foreland Basin

The North Pyrenean Thrust System consists of an imbricate system of north-verging thrust sheets (Muñoz 1992). Thrusting involved Hercynian basement and the Mesozoic and Tertiary cover (Fischer 1984, Déramond *et al.* 1993). The Aquitain retro-foreland basin evolved by flexure of the European plate in response to the load of the north Pyrenean thrust sheets (Desegaulx *et al.* 1990). During Tertiary compression, the thrust sheets were translated over the autochthonous parts of the Aquitain Foreland (Cámara and Klimowitz 1985, Williams 1985).

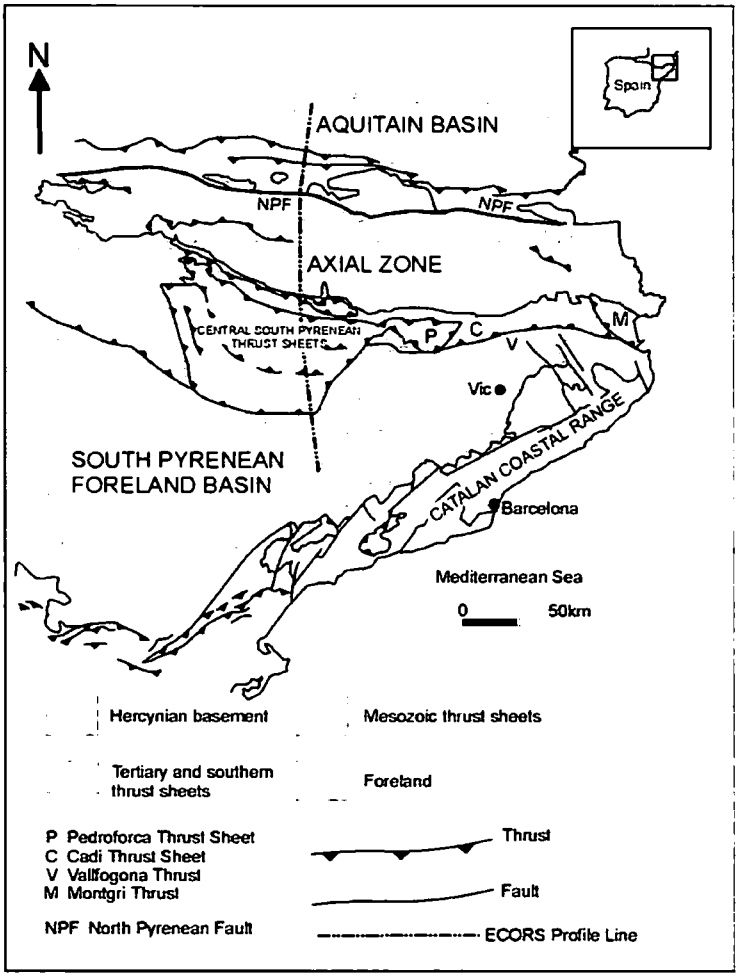


Figure 2.2 Tectonic domains of the Pyrenean Foreland Basin system. The Vic Basin is situated within the undeformed South Pyrenean Foreland Basin (also known as the Ebro Basin). After Puigdefabregas *et al.* (1986)

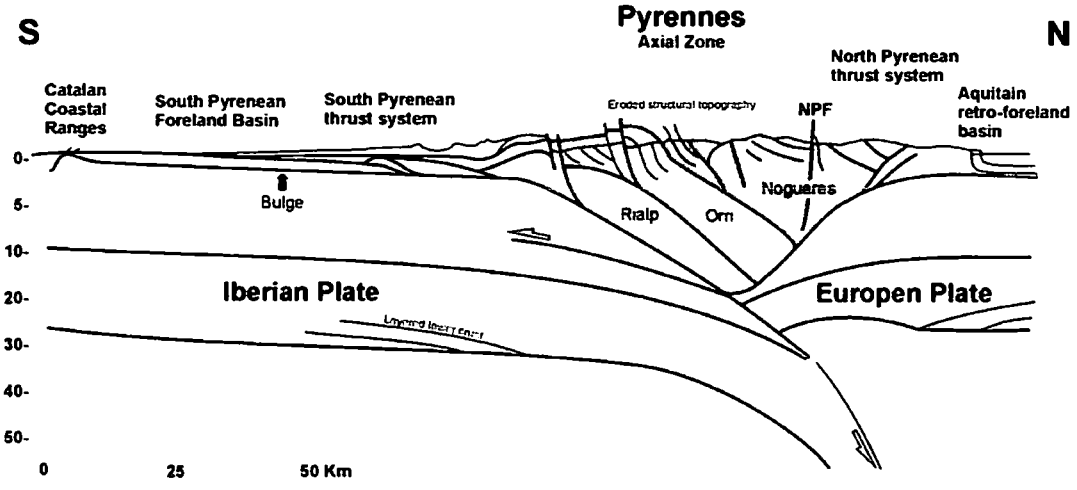


Figure 2.3 Summary sketch of the ECORS profile (the location of which is indicated on Figure 2.2), illustrating the structure of the Pyrenean Axial Zone. **NPF**=North Pyrenean Fault. After Verges *et al.* (1999).

2.3.2 The Pyrenean Axial Zone

The Pyrenean Axial Zone comprises the central part of the orogen (**Figure 2.2**), and consists of Cambrian to Carboniferous-age rocks, affected by low- to high-grade metamorphism during the Hercynian Orogeny and later intruded by granitic bodies in the Permian (Zwart 1979, Vissiers 1992). Post-Hercynian rocks uncomformably overlie the deformed basement (Vergés *et al.* 1995). The North Pyrenean Fault (NPF), bounding the northern part of the axial zone (**Figure 2.2**), is interpreted as the suture between the Iberian and European plates (Choukroue 1976, Puigdefàbregas and Souquet 1986, Muñoz 1992). Axial zone rocks are arranged in an antiformal stack of three main tectonic units named, from top to bottom, Nogueres, Orri and Rialp (Muñoz 1992). The present-day structure of the Axial Zone, determined from the results of the Pyrenees ECORS profile (Choukroune *et al.* 1989, Roure *et al.* 1989, Munoz 1992 and others) is illustrated in **Figure 2.3**. The incorporation of post-Hercynian rocks into these thrust sheets confirms that thrusting is Alpine in age (Muñoz 1992).

2.3.3 The South Pyrenean Thrust System and South Pyrenean Foreland Basin

The Southern Thrust System (**Figures 2.2** and **2.3**) extends laterally for over 1500 kilometres from east to west, measures up to 50 kilometres from north to south, and is considered the more significant of the two thrust systems (Puidefàbregas *et al.* 1986, Millán *et al.* 1995, Vergés *et al.* 1995). The Southern Thrust System comprises a piggyback sequence of three major thrust sheets known as, in order of emplacement, the Upper, Middle and Lower Thrust Sheets. The Upper Thrust Sheets were emplaced in the Late Cretaceous to Early Eocene and are composed of Mesozoic sedimentary rocks. The Middle Thrust Sheets were emplaced in the Eocene below the Upper Thrust Sheets. The sheets contain Hercynian basement rocks, a reduced Mesozoic cover and Paleogene foreland sedimentary sequences. The Lower Thrust Sheets were emplaced in the late Eocene-Oligocene and comprise Silurian basement and a much reduced Mesozoic cover section (Puigdefàbregas and Souquet 1986). Based on balanced cross-section data, the total estimated shortening across the eastern sector of the Southern Thrust System is approximately 146 kilometres or up to fifty percent (Puigdefàbregas *et al.* 1986, Vergés *et al.* 1995). Shortening rates decreased from a maximum of around 4.5 mm a^{-1} in the early- to mid-Eocene to around 2.0 mm a^{-1} by the end of Pyrenean deformation (Vergés *et al.* 1995).

The South Pyrenean Foreland Basin (**Figures 2.2** and **2.3**), within which the study area in NE Spain is situated, developed in the Tertiary as a direct result of the tectonic loading and southward displacement of the southern thrust sheets over a stable Lower Eocene carbonate platform. Overthrusting led to the fragmentation and geographic isolation of what were previously interconnected parts of the relatively depressed foreland (Puidefàbregas *et al.* 1986, Molenaar *et al.* 1996). The foreland basin developed on pre-Mesozoic basement and forms an irregular triangular shape bounded by the Pyrenees to the north and the Catalan Coastal Range to the southeast (Taberner *et al.* 1999). Sedimentation in the South Pyrenean Foreland Basin was contemporaneous with development of the thrust system (Puigdefabregas *et al.* 1986, Burbank *et al.* 1992). Palaeogene remnants of the foreland basin occur as allochthonous thrust sheets (**Figure 2.2**). The Iberian and Catalan Coastal Ranges, situated to the west and south-east respectively, bound autochthonous foreland strata (Puigdefabregas *et al.* 1986, Puigdefabregas and Souquet 1986, Burbank *et al.* 1992). Sediments along the northern margin of the Pyrenean Foreland are typically deformed, while sediments on the southern margin are predominantly undeformed.

2.3.4 The Catalan Coastal Range

The intraplate Catalan Coastal Range delimits the active distal margin of the Pyrenean Orogen and southeastern margin of the South Pyrenean Foreland Basin system (**Figures 2.2** and **2.3**). The Catalan Coastal Range is an important geological feature since it was a positive topographic feature in the Eocene shedding siliciclastic material to the southern foreland margin, and into the study area in the Vic Basin (Lopez-Blanco 1993, Monstad 2000). The Catalan Coastal Range affected the rates of subsidence across southeast foreland basin and therefore played an important role in the development of the basin stratigraphy (Vergés *et al.* 1998). The range developed as a major reactivated strike-slip fault system during the Alpine Orogeny 41.5 Ma ago (Anadon *et al.* 1985, Vergés *et al.* 1998). Structurally, the Catalan Coastal Range consists of two major late Palaeozoic and Mesozoic fault systems (Lopez-Blanco 1993). Alpine deformation reactivated these old fault lineaments within the Coastal Range as a series of NE-SW stepping, left lateral strike slip faults which extend laterally for about 200 kilometres (Lopez-Blanco 1993, Monstad 2000). A series of SE-NW oriented thrusts and folds were generated, uplifting slices of Lower

and Middle Triassic basement and Mesozoic cover as push-up blocks, folds and flexures forming a zone of high relief in front of the relatively depressed foreland (Monstad 2000).

A series of syntectonic alluvial and fan delta complexes (including the Montserrat and Sant Llorenç del Munt complexes) formed on the down-thrown side of the strike-slip fault system during the Lutetian to Priabonian (Lopez-Blanco 1993, Busquets 1995). Proximal fan conglomerates stacked up along the basin margins undergoing rapid uplift and erosion (Cabrera *et al.* 1986). Detailed studies of the conglomerates has yielded up to four orders of cyclicity due to relative sea level changes and variations in siliciclastic supply, both ultimately controlled by the tectonic situation at the basin margin (Lopez-Blanco 1993).

Alluvial fan systems evolved into fan deltas following a major transgression in the basal Bartonian (Lopez-Blanco 1993). In some areas, carbonate platform systems with an areal extent up to 12 km^2 developed. Carbonates are *Nummulites* shoals and bars, and barrier and coastal reef systems (Lopez-Blanco 1993, Monstad 2000).

The present day Catalan margin is characterised by Plio-Quaternary uplift, and a well-developed horst and graben structure extending to the Valencia Trough in the southeast (Millán *et al.* 1995).

2.4 The Southeastern Pyrenean Foreland Basin: Development and Stratigraphy

The latest Maastrichtian to early Oligocene sedimentary evolution of the Southeast Pyrenean Foreland Basin has been differentiated into nine depositional sequences in the sense of Vail *et al.* (1977) (**Figure 2.4**) (Puidefàbregas *et al.* 1986). "Each sequence comprises a body of genetically related strata bounded by unconformities, correlative conformities or by abrupt vertical changes in regional facies distribution" (Puidefàbregas *et al.* 1986). **Table 2.1** summarises important sedimentary formations within each depositional sequence. The Palaeogene South Pyrenean Foreland Basin-fill is broadly asymmetric in cross-section. The total sediment thickness is up to 4.0 km in the north, thinning to a few hundred metres in the south (Hirst and Nichols 1986). The most complete depositional record occurs in the thrust sheets to the north and along the southeast basin margin bordering the Catalan Coastal Range. In the centre and southeast, earlier depositional sequences are known only from well log data (Puidefàbregas *et al.* 1986).

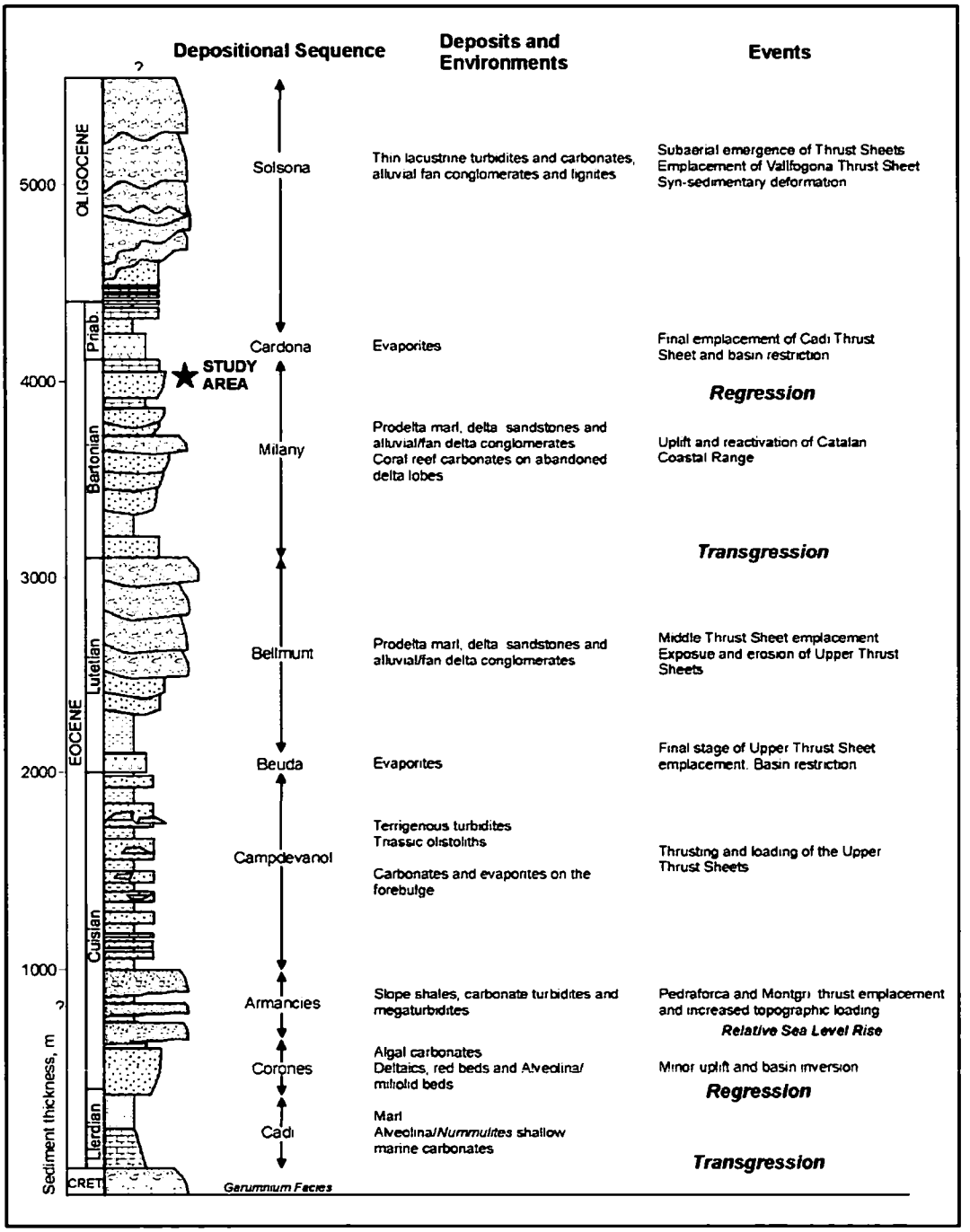


Figure 2.4 Summary stratigraphic log illustrating the nine tectono-srtatigraphic units of the Southeastern Pyrenean Foreland Basin. Note that all sequence thicknesses are estimated. The approximate stratigraphic location of the studied section is also indicated. Based on data from Puigdefàbregas *et al.* (1986)

Tectono-sedimentary Sequence	Units/Formations	Authors
Solsona	Artes Formation	Riba 1976 Busquets *et al.*, 1985
Cardona	Cardona Formation	Busquets *et al.*, I985 Taberner *et al.*, 1983 Puigdefàbregas *et al.*, 1975
Milany	Folgueroles Formation Collbàs Formation Igualada Formation Milany Formation Roca Corba Formation Santa Magdelana Formation Montserrat Formation Tossa Formation Collsuspina Formation	Regaunt 1967 Ferrer 1971 Ferrer 1971 Gich 1972 Palli 1972 Gich 1972 Anádon *et al.*, 1975 Ferrer 1971 Taberner *et al.*, 1999a Hendry *et al.*, 1999 Alvarez *et al.*, 1995
Bellmunt	Banyoles Formation Barcons Formation Bellmunt Formation Tavertet Formation Romgats Formation	Rios and Masachs 1953 Gich 1972 Gich 1972 Regaunt 1967 Colomba 1980
Beuda	Beuda Formation	Ortí *et al.*, 1985
Campdevànol	Turbidite sequences	Burbank and Puigdefàbregas 1985
Armàncies	Penya Formation Pontils Group	Estevez 1973 Anádon 1978
Corones	Sagnari Formation Corones Formation	Gich 1972
Cadí	Cadí Formation Orpí Formation	Mey *et al.*, 1968 Ferrer 1971 Gich 1972

Table 2.1 List of important sedimentary formations and authors classified into relevant depositional sequences in the Southeastern Pyrenean Foreland Basin.

2.4.1 Tectonosedimentary Units of the Southeast Pyrenean Foreland

Cadí Sequence

Shallow marine platform carbonates dominate the Cadí tectono-sedimentary sequence (Puidefàbregas *et al.* 1986, Burbank *et al.* 1992). The sequence represents the development of an extensive carbonate platform system on the continental red beds and lacustrine deposits of the Late Cretaceous Garumnian Facies (**Figure 2.4**) (Puidefàbregas *et al.* 1986 and references within).

Corones Sequence

Basin inversion in the Llerdian-early Cuisian occurred as a consequence of initial emplacement of the Upper Thrust Sheets (**Figures 2.4**). Deltaic sediments derived from the uplifted area in the north prograded southwards (Puidefàbregas *et al.* 1986 and references within). Minor tectonic uplift on the passive margin related to thrusting is inferred from the development of unconformities and deposition of southerly-derived siliciclastic material on the Llerdian carbonate platform (Burbank *et al.* 1992).

Armàncies Sequence

Rapid deepening in the foreland as a consequence of continued thrust sheet emplacement during the Cuisian is recorded in the sediments of the Armàncies sequence (**Figures 2.4**). This sequence also coincides with a high global sea level (Haq *et al.* 1987). The Armàncies sequence is characterised by slope deposits including shale, carbonate turbidites and megaturbidites. Within the megaturbidites are Llerdian-aged limestone clasts (derived from the carbonate platform of the Cadí sequence) within an upper-Cuisian *Nummulites*-rich matrix (Puidefàbregas *et al.* 1986). In summary, this tectono-sedimentary sequence represents the destabilisation of the carbonate shelf as a precursor to catastrophic shelf margin collapse during submarine thrusting. A contemporaneous carbonate platform developed over the southern passive margin as relative sea level rose.

Campdevànol Sequence

The tectonic load created as a result of Upper Thrust Sheet emplacement led to the formation of a deep turbidite trough in the late-Cuisian (**Figure 2.4**) (Burbank *et*

al. 1992). The turbidites of Campdevànol sequence indicate a sudden terrigenous influx and a switch from carbonate to siliciclastic deposition (Puidefàbregas *et al.* 1986, Burbank *et al.* 1992). Loading caused by emplacment of the Pedraforca and Montgri Upper Thrust Sheets also led to development of a fore-bulge along the southern passive margin of the foreland where carbonates and evaporates continued to develop.

Beuda Sequence

Eustatic Sea level fall of around 50 metres and basin restriction occurred at the Cuisian-Lutetian boundary (Haq *et al.* 1987). This coincided with final Upper Thrust Sheet emplacement and emergence, and deposition of evaporates and red beds of the Beuda Sequence (**Figure 2.4**) (Burbank *et al.* 1992).

Bellmunt Sequence

Middle Thrust Sheet (Cadí) emplacement commenced in the Lutetian (Burbank *et al.* 1992). This coincided with a rapid relative sea level rise (Haq *et al.* 1987). Erosion of the recently emergent Pedraforca thrust sheet fed alluvial fans and fan deltas of the Bellmunt Sequence, which prograded southward over the Beuda evaporites (**Figures 2.4**) (Martinez *et al.* 1988). The development of the Freser Valley antiformal stack below the Cadí thrust sheet (and consequent topographic rejuvenation) increased siliciclastic input and accelerated delta progradation (Puidefàbregas *et al.* 1986). Consequently, the carbonate platform on the southern basin margin retreated.

Milany Sequence

The transgressive base of the Bartonian succession in the southeast foreland shows a switch from alluvial to the shallow marine depositional systems of the Bartonian Milany Sequence (**Figure 2.4**) (Puidefàbregas *et al.* 1986, Santesteban and Taberner 1988). The Milany sequence is the final fully marine sedimentary succession in the southeastern foreland (Taberner *et al.* 1999). Sediments of the southerly-prograding delta systems of the Milany Sequence were sourced from granitic basement rocks brought to the surface through the out of sequence thrusting behind the Freser Valley antiformal stack. Reactivation of fault systems along the southern foreland margin, and consequent uplift and erosion of Triassic basement blocks, fed

numerous northward-prograding alluvial fan and fan delta systems (Anadon *et al.* 1985, Puigdefabregas *et al.* 1986, Lopez-Blanco 1993). High sedimentation rates, as a consequence of north and south progradation, contributed to the gradual restriction of the basin (Puigdefabregas *et al.* 1986). The Calders study area within the Vic Basin is situated within the northward prograding delta system of the southern foreland basin margin (**Section 2.5**).

Cardona Sequence

Deposition of evaporites of the Cardona depositional sequence coincide with the largest and most abrupt eustatic sea level fall of the late Eocene-Oligocene, and are associated with final basin restriction (Burbank *et al.* 1992). Puigdefàbregas *et al.* (1986) attribute the diminshing siliciclastic influx in the late Eocene to the fact that the Upper Thrust Sheets (including the Vallfogona thrust) had not yet emerged. The Vallfogona thrust sheet moved along a ramp and was finally emplaced in the Oligocene, defining the northern margin of the youngest South Pyrenean Foreland Basin (**Figure 2.4**) (Puigdefàbregas *et al.* 1996).

Solsona Sequence

The Solsona depositional sequence is the non-marine deposits that finally filled the basin (**Figure 2.4**) (Riba 1976, Busquets *et al.* 1985). Progressive unconformities developed within the continental conglomerates and red beds as a result of movement of the Vallfogona thrust sheet (Puigdefàbregas *et al.* 1986).

2.4.2 Controls on patterns of Pyrenean Foreland sedimentation

"Patterns of sedimentary infilling of a flexural basin respond to complex interactions of subsidence, sediment supply, base level change, sediment sorting and transport efficiency" (Burbank *et al.* 1992). Ultimately, basin geometry and patterns of sedimentation in the southeastern Pyrenean foreland were controlled by development of the thrust system. During the Pyrenean Orogeny, as deformation progressed southward, the northern parts of the foreland were progressively incorporated into the younger thrust sheets and basin depocentres migrated southward (Puigdefàbregas *et al.* 1986). **Figure 2.5** illustrates the massive depocentre migration between deposition of the Cadi to the Cardona Sequences. Periods of decreasing subsidence rates in the southern Pyrenean Foreland Basin are associated with the

progradation of coarse-grained facies. Conversely, periods of accelerating subsidence are times of fine-grained facies deposition according to Burbank *et al.* (1992). In addition, topographic relief produced by thrusting and the structural development of the neighbouring Catalan Coastal and Iberian Ranges, also played a fundamental role in the final sequence geometry and sediment supply, particularly to the southern margin of the southern foreland (e.g. Gimènez -Montsant and Salas 1997, Puigdefàbregas and Souquet 1986 and Guimerà 1984).

In summary, the late Cretaceous to early Oligocene history of the Pyrenean Foreland Basin can be broadly divided into two stages. The first, lasting from the late Cretaceous to the middle Lutetian (and consisting of the Cadi to the lower part of the Bellmunt depositional sequences of Puigdefabregas and Souquet (1986)) is characterised by low topographic relief, submarine emplacement of the thrust front, fast shortening rates, widespread marine deposition and an erosion rate virtually equal to detrital foreland accumulation. The second stage, lasting from the middle Lutetian to the Oligocene and consisting of the upper Bellmunt to Solsona depositional sequences, is characterised by increased topographic relief, sub-aerial emplacement of the thrust front, decreased shortening rates, widespread continental sedimentation and mountain erosion rates approximately three times basin accumulation.

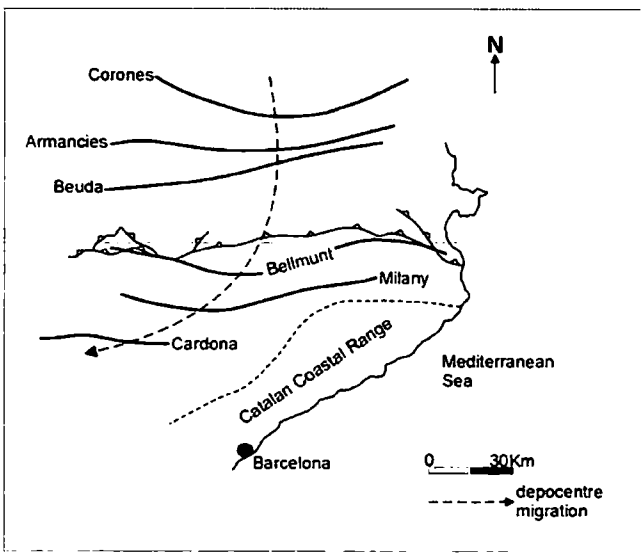


Figure 2.5 Schematic sketch illustrating the southerly migration of the Pyrenean Foreland Basin depocentre between deposition of the Corones and Cardona Sequences. The solid lines represent the southern extent of the depositional sequence. Modified from Puigdefàbregas *et al.* (1986)

2.5 Study Area: The Paleogene Vic Basin

The Vic Basin is located in the eastern-most Catalan sector of the Southeast Pyrenean Foreland Basin and is the only sector where both north and south-sourced Eocene marine platform depositional systems are exposed (Taberner *et al.* 1999). The sedimentary fill of the Vic Basin is Palaeocene to Oligocene in age, and the sedimentary succession is interpreted to represent final open marine sedimentation in the eastern foreland prior to widespread continental alluvial, lacustrine and evaporitic deposition (Taberner 1983, Taberner *et al.* 2000). The Vic Basin-fill spans the Bellmunt, Milany, Cardona and the Solsona Sequences illustrated on **Figure 2.4**. The basin deposits are largely undeformed and the present exposed margin of the basin is approximate to that of the Palaeogene margin (Taberner and Bosence 1985).

The Vic Basin existed in the Early Eocene as an east-west trending gulf that connected the Mediterranean to the Atlantic via the Navarra-Basque zone in the north-central Pyrenees (Alvarez *et al.* 1994). By the late Middle Eocene, the eastern connection to the Mediterranean through this gulf was severed. Connections with the Atlantic and Mediterranean remained open to the south (**Figure 2.6**). Thick marl successions were deposited in distal and deeper areas of the Vic Basin during the Middle Eocene, and north and south-sourced deltaic systems developed on the shallow northern and southern margins (**Figure 2.7**) (Alvarez *et al.* 1994).

The final stages of marine sedimentation in the Vic Basin during the Bartonian are known as the Marine Sequence (e.g. Taberner *et al.* 1999, Travé *et al.* 1996 and Burbank *et al.* 1992). The Montserrat conglomerates deposited as alluvial fan complexes on the northern margin of the Catalan Coastal Range (**Section 2.3.4**) interfinger with the Marine Sequence (Busquets 1995). The duration of the Marine Sequence was approximately 10 Myr, lasting from 46 Ma to 35 Ma (Taberner *et al.* 1999). Sediments of the Marine Sequence were successively deposited within increasingly southerly positions as the basin depocentre migrated from north to south as a consequence of thrust emplacement (**Section 2.4.2**) (Puigdefabregas *et al.* 1986). Strong basin-fill asymmetry is recognised during this time. Magnetozones in marginal marine sediments in the north are considerably thicker than those in the south. Continental and marine sediments derived from the Catalan Coastal Range onlap and eventually overlap the Palaeozoic and Mesozoic basement in an area known as the Centelles high, which is suggested to represent the southern Pyrenean fore-bulge

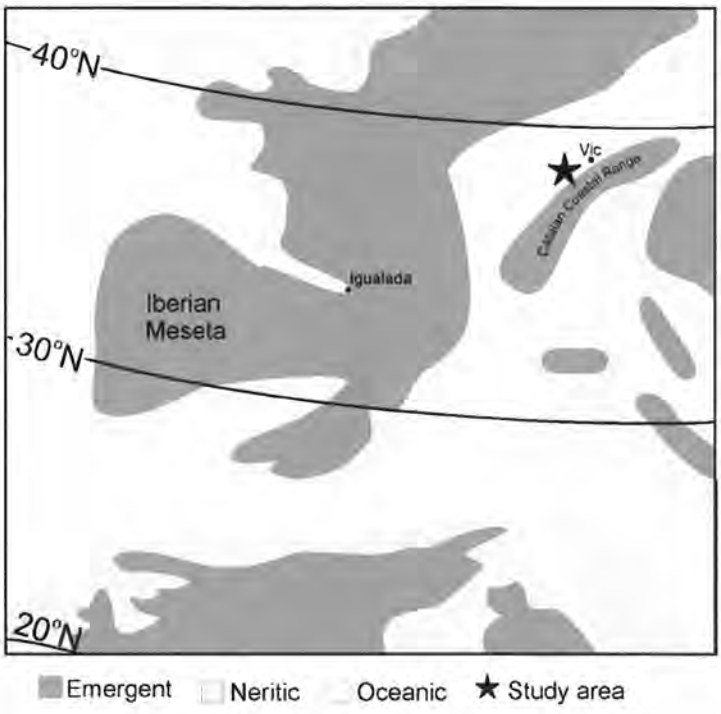


Figure 2.6 Palaoegeographic reconstruction of the Pyrenean region in the Middle Eocene. The studied section at Calders (indicated) was situated on the southern margin of the Vic Basin, and received siliciclastic detritus from the Catalan Coastal Range (indicated) that was a topographically positive feature in the Eocene. A tortuous route to the Atlantic Ocean and Mediterranean Sea remained open. Redrawn from Dercourt *et al.* (1985).

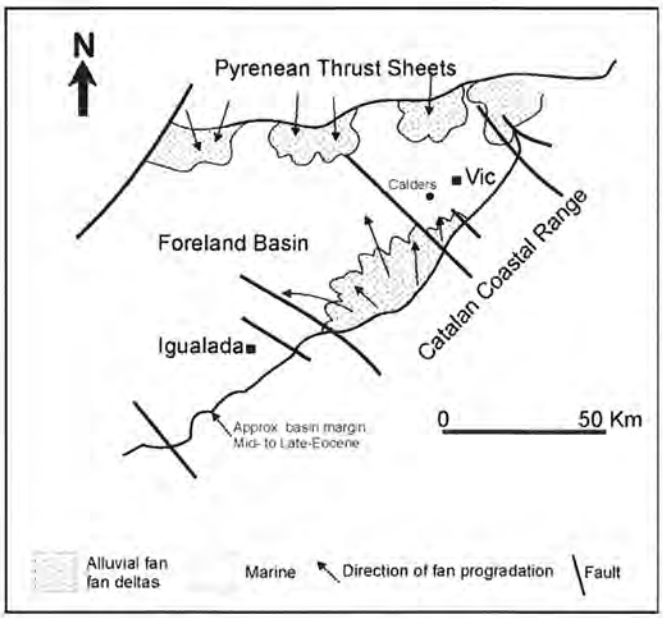


Figure 2.7 Mid- to Late Eocene palaeogeographic reconstruction of the Southeast Pyrenean Foreland Basin. Sediments in the northern areas were sourced from erosion of the axial zone thrust sheets and uplifted Hercynian granitic basement. Sediments in the southern areas, including the study area at Calders (indicated on the map), were sourced from the Mesozoic Catalan Coastal Ranges. Both coastal and marine systems on the northern and southern margins are classed at part of the Milany Sequence, although the two systems have distinctively different mineralogy that enables their distinction in the field.

(Taberner *et al.* 1999). For the duration of most of the Marine Sequence (up to 38.11 Ma) the Vic Basin therefore existed as two sub-basins, a northern and southern sub-basin, separated by the east-west trending Centelles high (Taberner *et al.* 1999). Time-equivalent units were deposited in each sub-basin, but similar facies types were not always contemporaneous on each margin (Taberner *et al.* 1999). Sediments in the southerly back-bulge area were derived from the southern margin of the basin (the tectonically re-activated Catalan Coastal Range), and sediments in the north were derived from erosion of the Pyrenean thrust sheets. The studied section at Calders was deposited in the southerly sub-basin.

The Middle Eocene marine and transitional sediments of the Vic Basin have been described as seven sedimentary cycles: four Lutetian cycles, two Bartonian cycles, and a final late Bartonian to early Priabonian cycle (Serra-Kiel *et al.* 1997). Each cycle is identified as the package of sediments deposited in the basin from the base of a transgressive system up to the base of the next transgressive system (Serra-Kiel *et al.* 1997, following Bates and Jackson 1987). The sedimentary cycles are summarised in the following sections.

2.5.1 Lutetian marine sedimentation in the Vic Basin

Lutetian marine sediments are only exposed in the eastern sector of the Vic Basin. Lutetian sediments overlie the laterally extensive fluvio-alluvial facies of the Pontils and Vilanova de Sau Formations (part of the Bellmunt Sequence). Total sediment thickness is approximately 1300 metres on the northern margin, thinning to only a few hundred metres in the south (Serra-Kiel *et al.* 1997). A summary of the Lutetian stratigraphy of the Vic Basin is illustrated in **Figure 2.8**.

2.5.1.1 Lutetian Cycles 1 and 2

The first two sedimentary cycles, within the Tavertet Limestone Formation, show deepening trends. Cycle 1 consists primarily of lagoonal deposits and Cycle 2 consists of siliciclastic nearshore facies with longshore bars and *Nummulites* banks. The top of the second cycle is an intensely bioturbated limestone rich in *Alveolina sp.*, *Orbitolites sp.*, miliolids, echinoids, oysters and ferruginous concretions, and is interpreted as a condensation layer by Serra-Kiel *et al.* (1997).

2.5.1.2 Lutetian Cycle 3

The third Lutetian sedimentary cycle is interpreted as a transgressive and regressive sedimentary package. Deposits at the base of the transgressive cycle onlap the Tavertet Limestone (**Figure 2.8**). The transgressive system is characterised by a 1000 metre thickness of blue marl and bioclastic sands that constitute the lower and middle parts of the Colla de Malla and Banyoles Formations. The marls interdigitate with marly limestone condensation horizons. These layers are rich in *Ophiomorpha* and *Thalassanoides* burrows, *Nummulites* and molluscs (Serra-Kiel *et al.* 1997). Towards the south, the marls of the third cycle pass laterally into the continental sediments of the Romagats Formation, which were deposited on the southern margin of the basin (Colombo 1980, Serra-Kiel *et al.* 1997).

The regressive system of Cycle 3 consists of the upper portion of the Colla de Malla and Banyoles Formations, and contains prodelta and distal platform sediments coarsening upward into the fluvio-alluvial facies of the Bracons and Bellmunt Formations (**Figure 2.8**). The thickness of the regressive portion of Cycle 3 is up to around 450 metres in the north, thinning and fining considerably to the south (Serra-Kiel *et al.* 1997).

2.5.1.3 Lutetian Cycle 4

The fourth Lutetian sedimentary cycle has an overall shallowing trend, although sediments on the northern and southern margins of the Vic Basin are considerably different at this time. On the southern margin, inner platform limestone beds consisting of *Nummulites* and *Alveolina* packstones pass laterally and vertically into the continental sediments of the Romagats Formation (**Figure 2.8**). The sediments of Cycle 4 were the most southerly deposited marine sediments in the mid-Eocene (Serra-Kiel *et al.* 1997). Siliciclastic input was more important on the northern margin during deposition of Cycle 4, and limestones are relatively rare.

2.5.2 Bartonian marine sedimentation in the Vic Basin

Bartonian marine sediments of the Catalan sector of the Southeast Pyrenean Foreland are exposed in the Vic and Igualada Basins (Serra-Kiel *et al.* 1997). In both basins, the Bartonian is represented by two transgressive-regressive cycles composed of siliciclastics and carbonates (Romero *et al.* 2001). A summary of Bartonian stratigraphy is illustrated in **Figure 2.9**.

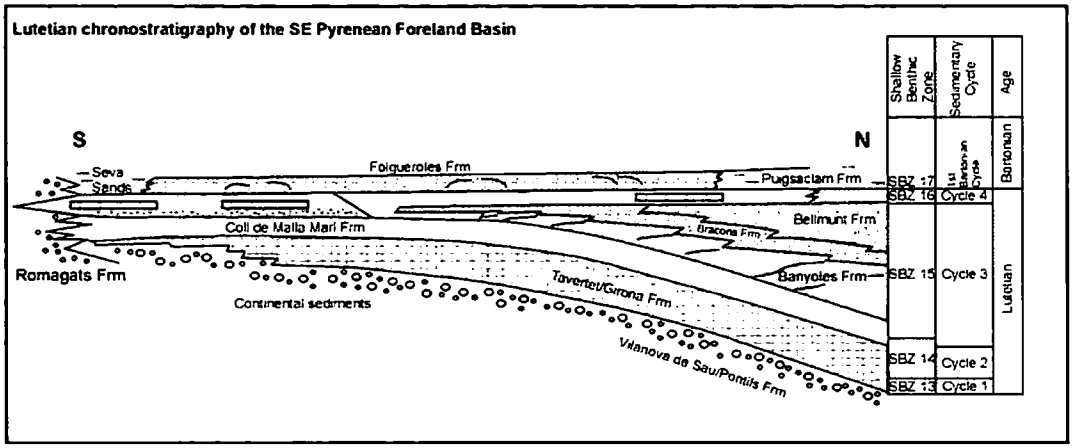


Figure 2.8 Lutetian chronostratigraphy of the SE Pyrenean Foreland Basin. Note the strong basin assymetry. After Serra-Kiel *et al.* (1997)

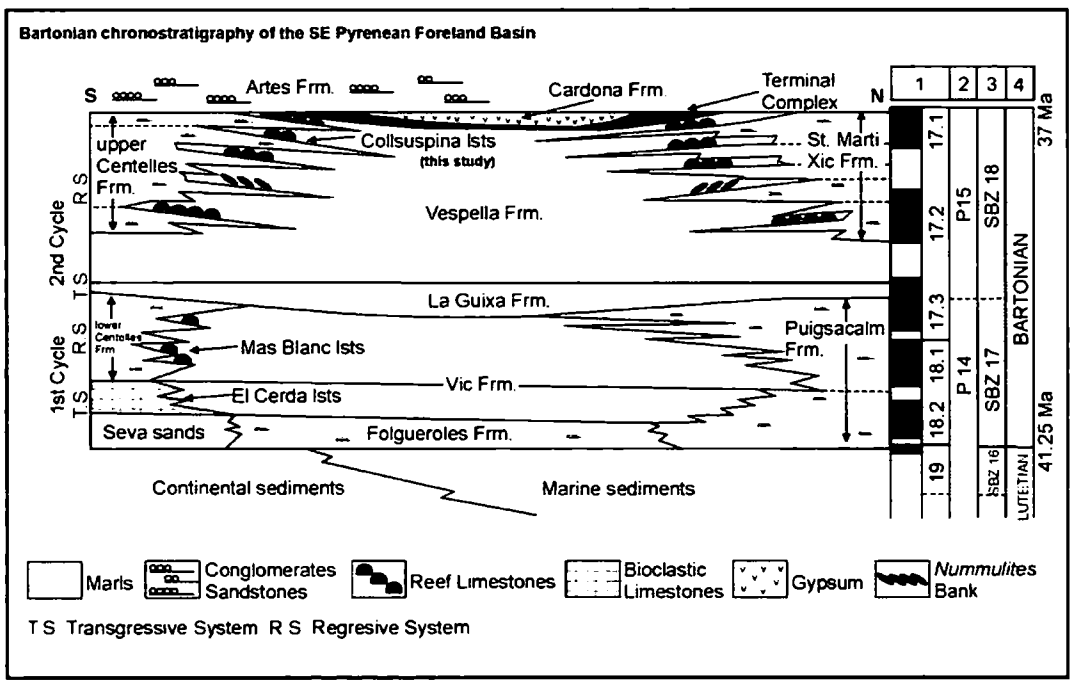


Figure 2.9 Bartonian chronostratigraphy of the SE Pyrenean Foreland Basin. The studied succession exposed at Calders is part of the Collsuspina limestones (indicated), located within the regressive part of the second Bartonian cycle. After Serra-Kiel *et al.* (1997)

2.5.2.1 Bartonian Cycle 1

At 46 Ma, an isochronous transgression affected the Vic Basin. The subsequent depositional conditions were open marine (Taberner *et al.* 1999). The transgressive beach facies of the Seva Sandstones and the El Cerdà Limestone Formation overlie continental late-Lutetian sediments on the southern basin margin (**Figure 2.9**) (Taberner 1983). The sandwave system of the glauconitic Folgueroles Formation overlies Lutetian marine sediments in the basin-centre, with the deltaic Puigsacalm Formation on the northern margin (Gich 1969, Reguant 1967, Taberner *et al.* 1999). The larger-foraminifera (*Nummulites, Operculina, Discocyclina*) rich marls of the Vic Formation in the basin centre, complete the transgressive part of this cycle (Reguant 1967, Serra-Kiel and Travé 1995, Serra-Kiel *et al.* 1997).

On the southern basin margin, the regressive system of the first Bartonian cycle consists of prograding siliciclastic sediments of the lower part of the Centelles Formation that alternate with carbonate buildups of the El Mas Blanc Limestone Formation (**Figure 2.9**) (Serra-Kiel *et al.* 1997). On the northern margin, the regressive system consists of prograding sandstones of the Puigsacalm Formation that change basinward into the upper part of the Vic Marl Formation (Serra-Kiel *et al.* 1997).

2.5.2.2 Bartonian Cycle 2

The transgressive system of the lower part of the second Bartonian cycle is represented by a siliceous sponge rich marl in the centre of the Vic Basin (the Guixa Marl Formation of Reguant (1967)). The marl becomes rich in bryozoa and larger foraminifera towards the basin margins (Serra-Kiel *et al.* 1997). The top of the transgressive system consists of the Gurb Marl Formation, which is rich in bryozoa, siliceous sponges, larger foraminifera, gastropods and decapods (Reguant 1967).

The northern and southern margins of the regressive system of the second Bartonian cycle are characterised by delta-reef complexes (Taberner and Bosence 1985, Santisteban and Taberner 1988, Alvarez *et al.* 1994, Hendry *et al.* 1999). The reef buildups of the Collsuspina Formation alternate with the sandstones of the upper part of the Centelles Formation on the southern basin margin. The Calders study area (**Section 2.5.3**) is situated within this regressive cycle and its stratigraphic location is indicated on **Figure 2.9**. Deltaics and reef systems of the Sant Martí Xic Formation on

the northern basin margin pass basinward into the non-fossiliferous marl of the Vespella Formation (Serra-Kiel *et al.* 1997 and Reguant 1967).

2.5.3 Bartonian-Priabonian sedimentation in the Vic Basin

The Bartonian and Priabonian sediments of the Vic Basin record the very final phase of marine sedimentation in the Pyrenean foreland, and the transition to continental sedimentation as the foreland switched from the underfilled to overfilled stage of its development (in the sense of DeCelles and Giles 1996).

2.5.3.1 The Terminal Complex

Defined by Travé (1992), the Terminal Complex is "an interval of sediments overlying the Milany Depositional Sequence, bounded at the top and bottom by two regional unconformities, exposed along the entire eastern margin of Southeast Pyrenean Foreland Basin". The Terminal Complex, accompanied by the Cardona Evaporites, is the final interval of marine sedimentation in the Southeastern Pyrenean Foreland Basin (Travé *et al.* 1996). Sedimentological analysis of a small portion of the Terminal Complex in the Sant Amanc area (immediately to the east of the Calders area) has formed a part of this study (**Section 2.5.5**).

The terminal complex facies shoal upward and accumulated during a gradual change from mesotrophic to eutrophic conditions as basin restriction commenced (Travé *et al.* 1996). The complex can be divided into three lower siliciclastic horizons and three upper carbonate units (Serra-Kiel *et al.* 1997 and Travé *et al.* 1996). The Lower Cabonate Unit was deposited in normal marine salinity in a mesotrophic platform environment colonised by foraminifera and coralline algae. The Middle Carbonate Unit, characterised by coralline algae bioherms alternated with cyanobacterial mats, was deposited in a platform environment controlled more by local variations in nutrient and organic resources, salinity and freshwater influx. The Upper Carbonate Unit is characterised by an abundance of cyanobacterial mats indicating more eutrophic conditions and a greater meteoric influence than in previous units (Travé *et al.* 1996).

2.5.3.2 The Cardona Evaporites

Evaporite deposition in the Vic Basin commenced 35 Ma and continued through to the Priabonian (Taberner *et al.* 1999). During the earliest stages of evaporitic deposition in the basin centre, the basin margins were areas of oolitic and stromatolitic development (Travé 1992, Hendry *et al.* 1999, Taberner *et al.* 1999). The sedimentary record of the Terminal Complex is considerably thicker in the north than the south of the Vic Basin, suggesting a diachronous basin transition (Taberner *et al.* 1999).

The evaporites of the Cardona Formation cover an estimated area of 700 km^2, and can be observed to in-fill marine depocentres after normal marine sedimentation ceased (Taberner *et al.* 2000). Carbonate and sulphate deposits outcrop along the Paleogene basin margins, and evaporitic facies (i.e. anhydrite, halite etc.) occupy the relatively deeper areas. The evaporitic sequence comprises in ascending stratigraphic order 1) laminated and selenitic gypsum pseudomorphs, 2) banded halite, 3) a potash unit and 4) alternating cycles of halite, clay and thin laminated anhydrite (Taberner *et al.* 2000). The potash unit was precipitated when the basin was virtually isolated from the open sea, and the upper halite unit is representative of a continental regime (Taberner *et al.* 2000).

2.5.4 Lateral correlation with the Igualada Basin

Bartonian sedimentary cyclicity is also recognised in the Igualada Basin, which is located to the southwest of the Vic Basin (**Figures 2.6** and **2.10**). The first Bartonian transgressive-regressive cycle in the Igualada Basin is referred to as the Collbàs Cycle, the second as the Igualada Cycle (Serra-Kiel *et al.* 1997, Romero *et al.* 2002). The stratigraphy of the Igualada Basin is dominated by marls and reefal carbonates (the Igualada Marl Formation and the Collbàs and Tossa Formations respectively), and there is a markedly less-significant siliciclastic influence compared to the Vic Basin as marls and reefal limestones dominate (Romero and Caus 2001). The lateral equivalent of the Collsuspina Formation of the Vic Basin is the Tossa Formation in the Igualada Basin (Serra-Kiel *et al.* 1997). These formations are time equivalent, but are isolated systems (**Figure 2.10**). The late-Bartonian to Priabonian sediments of the Igualada Basin is also represented by the Terminal Complex and Cardona Evaporites which precede final basin emergence and infilling by fluvio-alluvial sediments of the Artès Formation (Serra-Kiel *et al.* 1997).

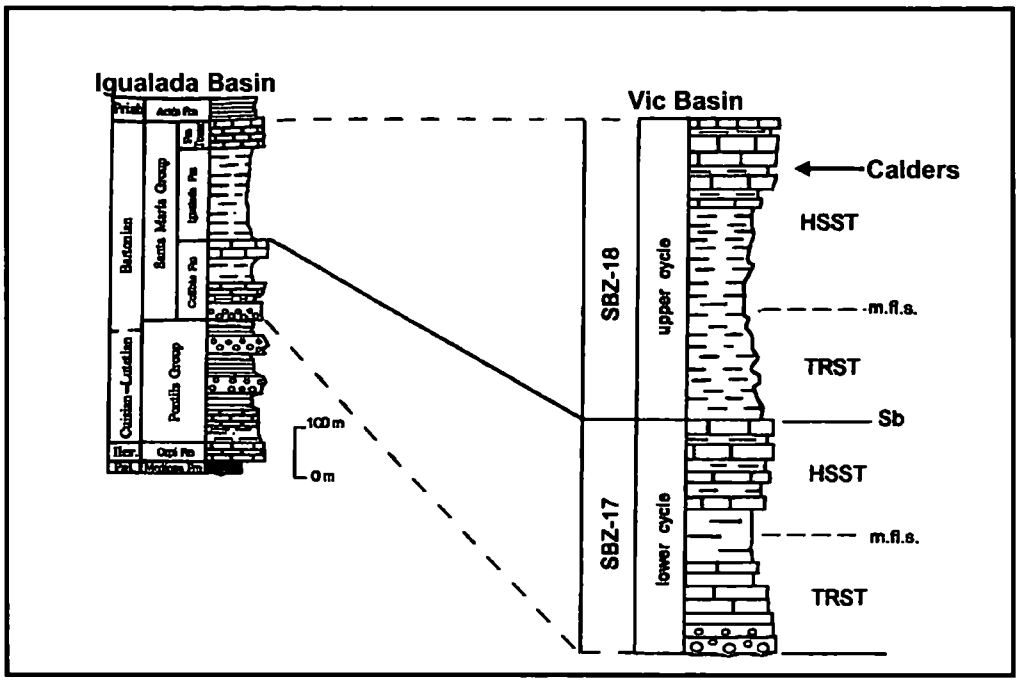


Figure 2.10 Correlation of Bartonian sedimentary cycles of the Igualada and Vic Basins. The Calders study area (indicated) in situated towards the top of the second Bartonian cycle in the Vic Basin. **TRST**= transgresive systems tract **HSST**=highstand systems tract, **m.fl.s.**=maximum flooding surface, **Sb**=sequence boundary. **SBZ**=Shallow benthic zone. The Calders study area is situated within SBZ 18, characterized by the larger foraminifera *Nummulites bedai.* Modified from Romero and Caus (2002).

2.5.5 Study Area

The Calders and Sant Amanc study area is located on **Figure 2.11**. The study area is covered by the 1: 50 000 Manresa topographic map (Sheet 363) published by the Instituto Tecnológico Geominero de Espana (ITGE). In addition, the 1:20 000 Moia Sheet 363-II, also published by the ITGE, was used.

2.5.5.1 Calders and Sant Amanc

The Calders section is situated immediately south of the village of Calders, approximately half way between Manresa and Moià (**Figures 2.11** and **2.12**). Exposure is as roadcut sections along the B 124 road running from Calders to Monistrol de Calders between K36 and K38.5, and within adjacent small east-west trending valley systems (**Figure 2.12**). The area of the Calders section is approximately 4.0 km^2. The stratigraphic thickness is 130 to 135 m. Calders is the

type locality of *Nummulites bedai*, which is characteristic of Shallow Benthic Zone 18 of Serra-Kiel *et al.* (1998). The Sant Amanc section is located in the far northeastern margin of the Calders area (**Figure 2.12**). The studied succession is situated within the regressive system of the second Bartonian cycle (**Figure 2.9**) and was deposited on the southern basin margin (**Figures 2.6** and **2.7**).

Lithological units exposed in the Calders section are (from base to top), the Centelles Formation (including the Collsuspina limestones), the Complejos de Calders Formation and the Artès Formation (after the terminology used by Hendry *et al.* 1999). The lithological units of the Calders section are part of the Milany and Solsona depositional sequences of Puigdefabregas *et al.* (1986). The Sant Amanc section is an exposure of the Teminal Complex (**Section 2.5.3.1**). Also exposed are the Complejos de Calders Formation and the Artès Formation.

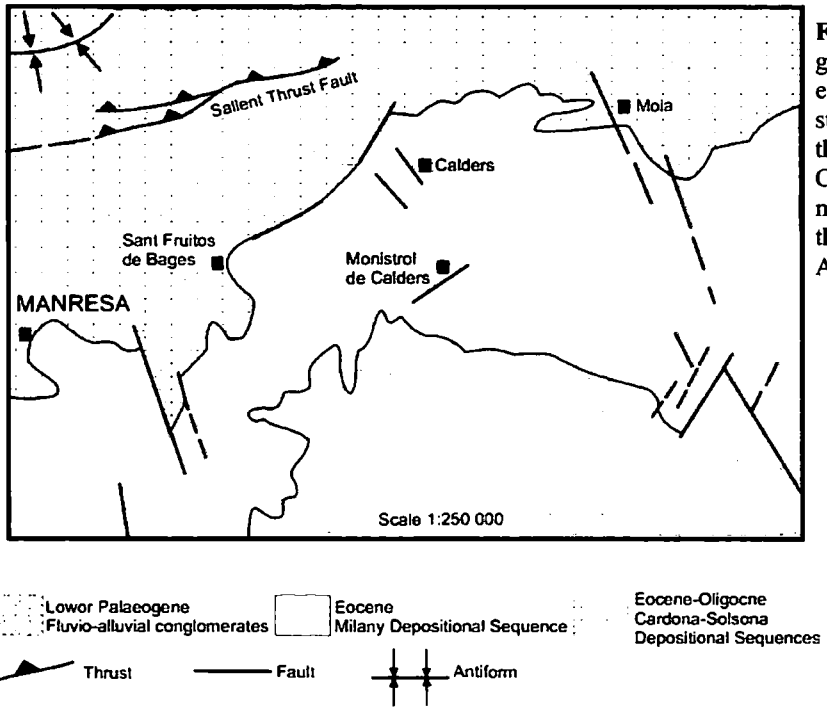


Figure 2.11 Summary geoglogical map of the eastern Vic Basin. The study area is situated to the immediate south of Calders (indicated on map). The sediments in this area are undeformed. After Peon *et al.* (1973)

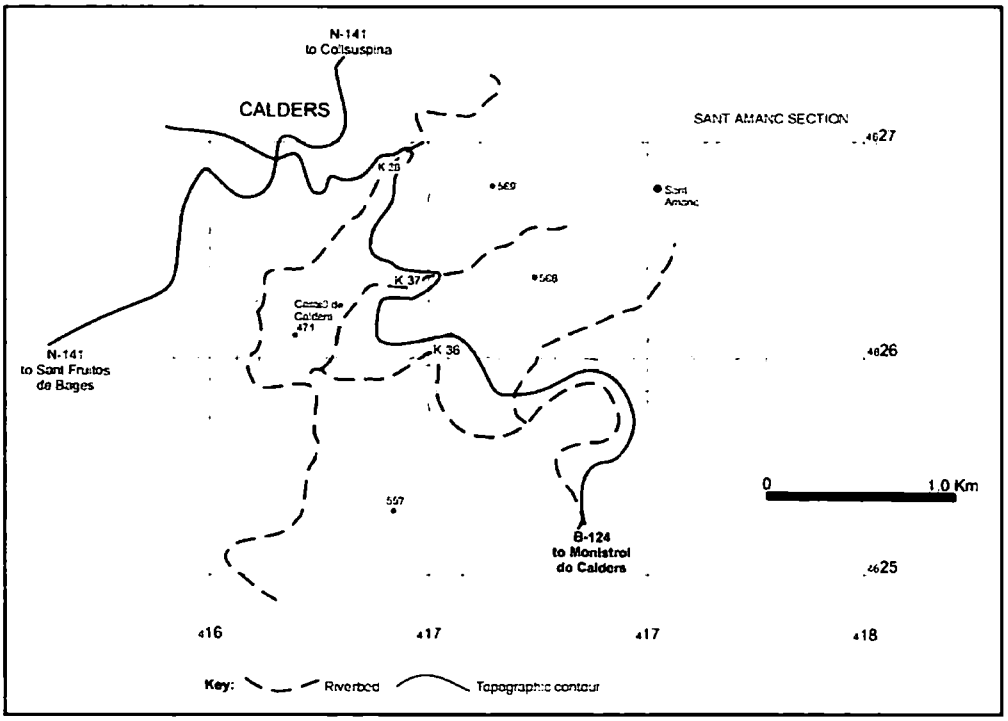


Figure 2.12 Summary topographical map of the study area. The Centelles Formation and the intercalated Collususpina Limestones are continuously exposed along the B124 road that runs between Calders and Monistrol de Calders (indicated). Further exposures are identified within the NE-SW trending valleys.

3. Calders facies and palaeoenvironmental analysis

3.1 Introduction

The principal aims of Chapter 3 are to present and synthesize facies data from the Calders and Sant Amanc areas, and to present a three-dimensional sedimentological/palaeoecological depositional reconstruction. A summary of the regional context of the studied sections and a description of sediments underlying and overlying the Calders section is presented in **Section 3.2**. **Sections 3.3** and **3.5** describe and interpret the different facies identified within the Calders and Sant Amanc successions respectively. Facies associations and three-dimensional palaeoenvironmental evolutionary models are proposed in **Sections 3.4** and **3.6**. It has not been the aim of this thesis to study the diagenesis of these sediments. However, a summary of the main diagenetic phases present is included within the facies descriptions. For a detailed review of the diagenesis of the Calders carbonates, refer to Hendry *et al.* (1999).

The Tertiary units of the Southeastern Pyrenean Foreland system have been described in detail by many authors, particularly in terms of sedimentary cyclicity in response to the eustatic and tectonic evolution of the foreland system (Taberner *et al.*, 1999, Burbank *et al.*, 1992, Puigdefàbregas and Souquet 1986, Puigdefàbregas *et al.*, 1986). The succession has also been studied with regards to the apparent juxtaposition of shallow marine carbonate and siliciclastic facies (Hendry *et al.*, 1999, Taberner and Bosence 1995, Santisteban and Taberner 1988). For a complete review of the stratigraphy and sedimentology of the Southeastern Pyrenean Foreland, refer to **Chapter 2**. This chapter presents the first complete facies scheme for the Calders area of the Vic Basin based on outcrop, petrological and micropalaeontological data. A summary map of the studied area illustrating the locations of logged sections and collected samples is presented on **Figure 3.1**. The logging, sampling and petrological techniques used are described in detail in **Appendix 1**. Thin-section data is presented in **Appendix 3**. **Appendix 5** presents logged sections that encompass field and thin section information. Samples have been classified using the limestone classification schemes of Dunham (1962) and Insalaco (1998), and are summarised in **Appendix 1**. The identification of larger benthic foraminifera assemblages has aided environmental interpretation of facies. For detailed information on the palaeoecology of larger benthic foraminifera, refer to **Appendix 2**.

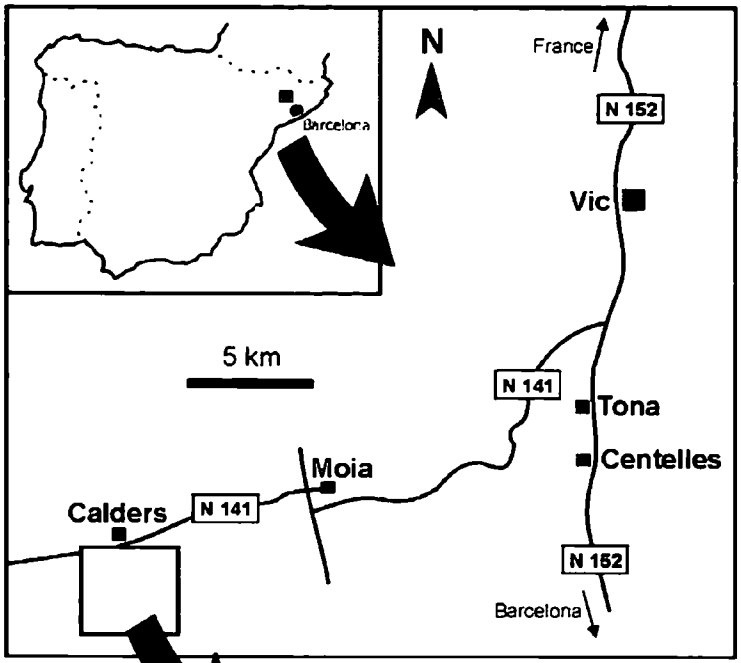


Figure 3.1a Regional location map of the study area.

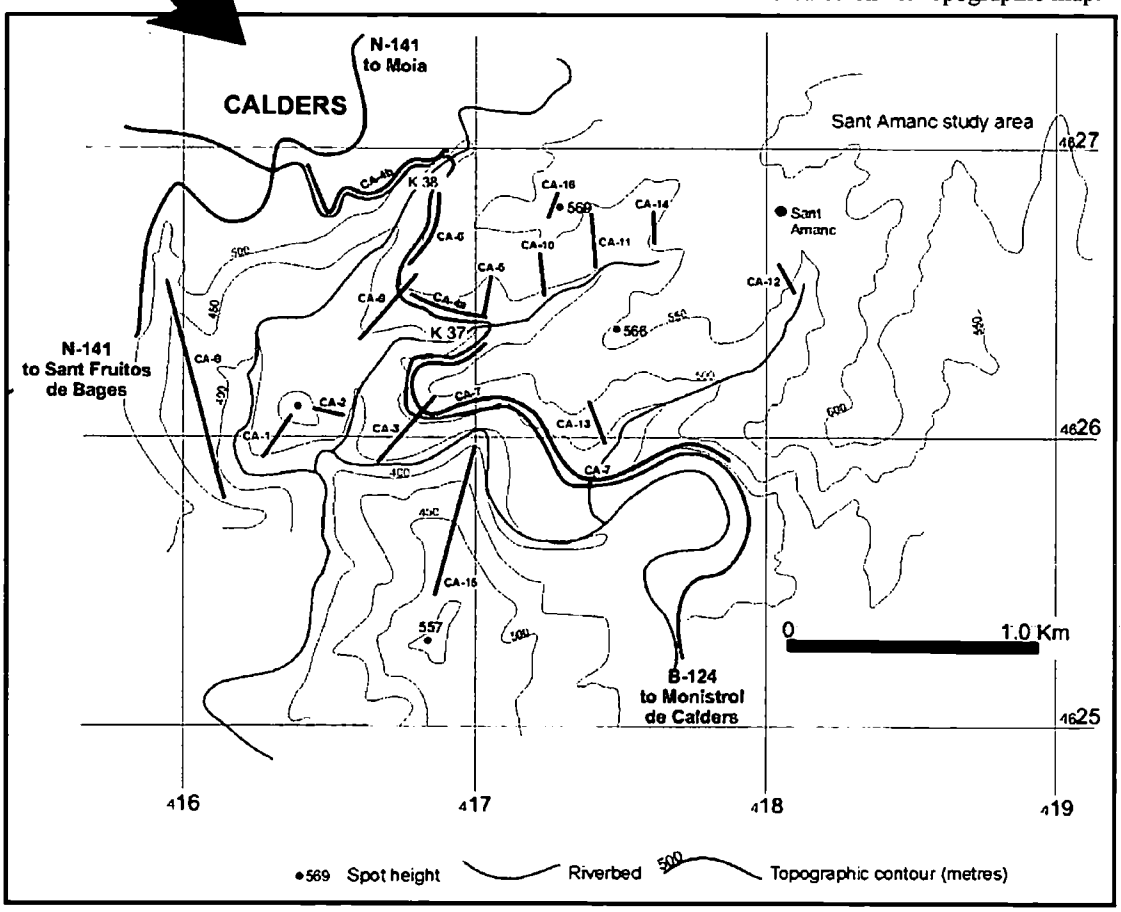


Figure 3.1b Summary topographic map of the study area at Calders. The location of logged sections is indicated on the topographic map.

Depositional sequence and Facies	Dominant textures	Bed thickness	Bedding characteristics	Bed contacts	Lateral continuity	Components	Clastic %	Sedimentary structures	Energy level	Environmental interpretation
Solsona Sequence										
Red mudstone (Artès Formation) **Section 3.2.5**	Mudstone Arenite	Unconstrained	Weathering of mudstones.	Sharp, non-erosive	> 4km (inferred)	Well-sorted quartz within red mud matrix	100 %	Lamination	Low	Continental, possibly alluvial
Cross-stratified gravelly litharenite (Complejos de Calders Formation) **Section 3.2.2**	Conglomerate Pebbly mdstone	0.25 to 0.5 m	Coarse lithic grains weather out of mudy matrix.Rare bioturbation.	Sharp, non-erosive	> 4 km (inferred)	Well-rounded lithics within muddy matrix. Non-fossiliferous.	Up to 100 %	Trough cross beds, planar cross beds (dm)	Mod. to high	Fluvio-deltaic
Milany Sequence (Terminal Complex?)										
Miliolid micritic arenite **Section 3.2.3**	Muddy arenite with bedding-parallel reddened horizons	0.9 to 1 m	Apparently tabular units. Bioturbation.	Sharp, non-erosive	Cannot be determined (lack of continuous exposure	Quartz, lithics, K-feldspar, miliolids.	Up to 50 %	Massive	Low to mod.	Low-energy marine, partially restricted.
Pelletal wacke/grainstone **Section 3.2.4**	Grainstone. Locally wackestone	Up to 1 m	Apparently tabular	Sharp, non-erosive	Cannot be determined (lack of continuous exposure	Faecal pellets, micrite	< 5% (clays)	Massive	Mod.	Restricted marine
Milany Sequence										
Arenaceous marl (Vespella Formation) **Section 3.2.1**	Marly sandstone Quartz wacke	10 cm to 1m	Apparently tabular units. Muddy weathering.	Poorly defined. Gradational contacts.	Cannot be determined (lack of continuous exposure).	Molluscs.	30 to 50 %	Diffuse lamination (mm)	Very low	Open marine, below fwwb. Distal delta front.

Table 3.1 Facies scheme for sediments that occur stratigraphically above and below the main Calders and Sant Amanc sections.

3.2 Regional context and associated sediments

Formations present in the study area, from oldest to youngest, are Vespella Formation, Centelles Formation (including the Terminal Complex), Complejos de Calders Formation and the Artès Formation. The Centelles Formation and the Terminal Complex formed the focus of study for this thesis, thus less attention has been paid to the other formations exposed. However, in order to clarify the stratigraphic and palaeoenvironmental context of the studied section, the facies encountered in the sediments underlying and overlying the main Calders reef are described in the following sections. The facies scheme for these sediments is presented in **Table 3.1**.

3.2.1 Arenaceous marl

(Vespella Formation, Milany Depositional Sequence)

Lithologies: Marly sandstone

Occurrence and bed characteristics

Extensive thicknesses of this facies are encountered 2.5 km to the south of Calders village between K 35 and K 34 along the B124 road (**Figures 3.1** and **3.2a**). This facies is characterised by a succession of homogeneous diffusely laminated marly sandstones. Bed thickness ranges from 10 cm to 1 m, although bedding is poorly defined (**Figure 3.2**). The lateral extent of this facies cannot be determined due to the lack of laterally continuous exposure.

Lithological description

The arenaceous marl facies is poorly consolidated. Exposed surfaces weather to a light brown to pale-grey colour (**Figure 3.2b**). Fresh surfaces are a blue/grey colour. This facies is very well sorted, comprising 50 to 70 % micrite matrix. Detrital siliciclastic grains, comprising up to 50 % of this facies, are sub-spherical to sub-angular. Grain size ranges from 0.1 to 1 mm. Siliciclastic grains are monocrystalline quartz, lithics, orthoclase feldspar and calcite. Body fossils comprise up to 10 % of this facies. Fossils include gastropods up to 4 cm in length, with oysters, pectinid bivalves and undifferentiated skeletal debris.

Figure 3.2a Outcrop exposure of the marly litharenite (part of the Vespella Formation). The height of this exposure is approximately 30 metres. Photograph taken from GR 18252530, facing NW.

Figure 3.2b Poorly-defined lamination in the marly litharenite facies. Photograph taken from GR 18292530, facing NW. Hammer=45cm.

Interpretation: depositional environment

A marine depositional environment is concluded for the arenaceous marl facies from the presence and moderate abundance of marine fauna. Deposition at depths below the lower limit of the photic zone is suggested from the absence of photosynthetic organisms in this facies. Deposition from suspension in a low-energy setting below fair weather wave base is inferred from the high-percentage of fine-grained material and the presence of fine laminations. Deposition some distance from the sediment source is concluded from the textural and compositional maturity of detrital siliciclastic grains. Due to the association of the arenaceous marl with the overlying Centelles Formation (**Section 3.3**), a relatively low-energy deltaic environment of deposition (distal delta front) is inferred.

3.2.2 Cross-stratified gravelly litharenite

Complejos de Calders Formation (Solsona Depositional Sequence)

Lithologies: Pebbly litharenite
Pebbly litharenitic siltstone

Occurrence and bed characteristics

This facies is encountered in isolated outcrops along the N 141 approximately 1 km to the west of Calders (**Figure 3.1**). Isolated outcrops of this facies are also exposed in the Sant Amanc study area overlying the Terminal Complex (**Section 3.6**) thus this facies extends laterally for at least 4 km. Measured bed thickness is 0.25 to 0.5 m. This facies is characterised by decimetre-scale trough cross-stratified pebbly litharenitic sandstone that grades vertically into less pebbbly, planar cross-stratified sandstone (**Figures 3.3a and c**). Trough cross beds range from 5 to 10 cm in height. Planar cross-sets are up to 15 cm in height and 1.5 m in length (**Figures 3.3b and c**).

Lithological description

The cross-stratified gravelly litharenite facies is poorly consolidated, and the fine matrix has preferentially weathered out. Exposed surfaces weather to a light brown colour (**Figure 3.3**). Fresh surfaces are a light brown-buff colour. This facies is matrix-supported and very poorly sorted (**Figures 3.4b and c**). Grain size ranges from 0.01 mm to 0.5 cm. Mean grain size is around 1.75 mm. Siliciclastic grains are unstrained and strained quartz, chert, lithics, orthoclase feldspar and calcite. Monocrystalline and polycrystalline quartz comprise up to 45 % of grains within this

Figure 3.3 a) to c) Field exposures of the cross-stratified gravelly litharenite facies situated to the west of Calders along the N 141 (GR 15402619). Scale bar=2m. **b)** Trough-cross stratified beds (i) overlie a nodular, poorly fossiliferous carbonate unit thought to be a component of the Terminal Complex exposed in the Sant Amanc area. Trough-cross beds pass vertically in planar cross-stratified pebbly mudrocks (ii). Scale bar=50cm. **c)** Cross-stratification is defined by alignment of pebbly lithic clasts (blue arrow). Scale bar=15cm.

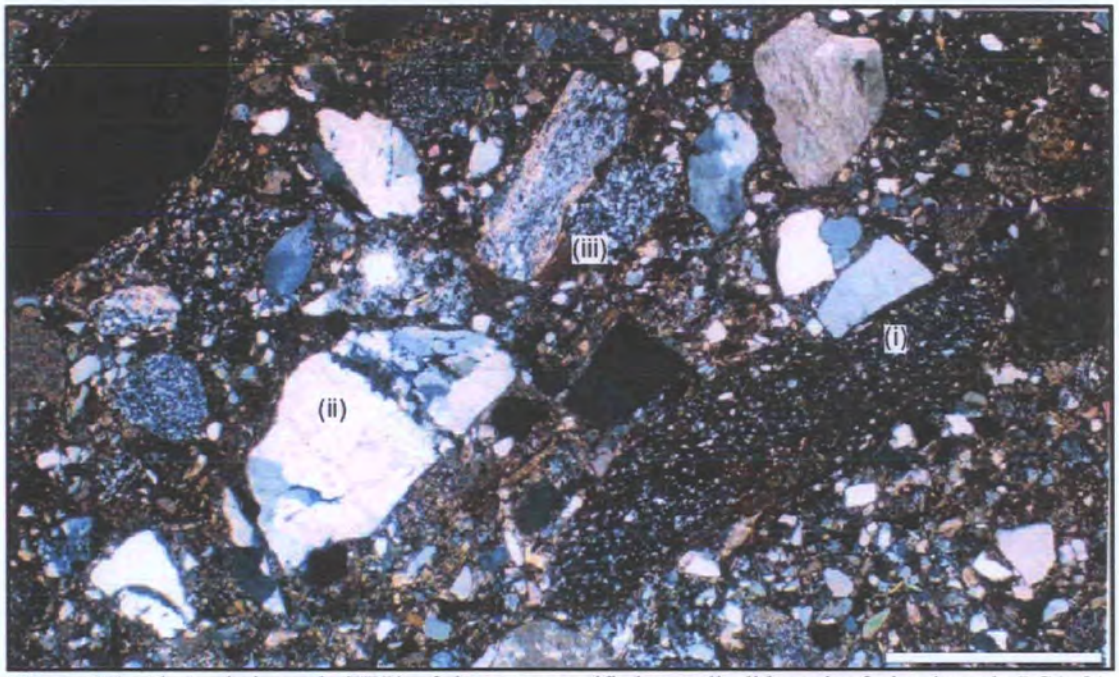


Figure 3.4a Photomicriograph (XPL) of the cross-stratified gravelly litharenite facies (sample LCA 81, GR 15402619). (i) lithic grain (siltstone), (ii) silty sandstone, (iii) chert. Coarse grains float within a muddy matrix. Scale bar=2 mm.

Figure 3.4b *Thalassanoides* burrows (arrowed) within a muddy horizon of the cross-stratified gravelly litharenite facies. Sant Amanc area. Scale bar= 50 cm.

facies. Quartz grains are typically angular with a low sphericity. Lithic grains compose up to 45 % of grains (siltstone with minor quartz-mica schist). Lithic grains are typically very well rounded with moderate sphericity and micas have commonly broken-down into yellowish clay observed in thin section. Angular calcite grains up to 0.25 mm comprise less than 1 % of grains within this facies. The matrix of the cross-stratified gravelly litharenite facies comprises mixed carbonate siliciclastic mud to fine sand. Matrix comprises up to 50 % of this facies.

Body fossils are absent in the cross-stratified gravelly litharenite facies. Peloids up to 2 mm are relatively rare (less than 1 % of grains). *Fugichnia* and *Thalassanoides* trace fossils are common in the upper, less pebbly parts of the cross-stratified gravely litharenite facies (**Figure 3.4b**).

Interpretation: depositional environment

A high energy, marine depositional environment above fair weather wave base is inferred for the cross-stratified gravelly litharenite facies due to the presence of vertical burrows and trough cross-stratification. A proximal depositional environment is inferred from the compositional immaturity of this facies. High sedimentation rates are inferred from the presence of fugichnial escape burrows. Substrate instability due to high sedimentation rates is interpreted to have inhibited colonisation by calcareous benthic organisms. However, waning energy is inferred from the gradual change from trough to planar cross stratification and the gradual fining upward within this facies.

In summary, deposition within a high-energy, fluvial-marine environment with high sedimentation rates is interpreted for the cross-stratified gravely litharenite facies. Hendry *et al.* (1999) have interpreted this facies as fluvio-deltaic. This facies is part of the Complejos de Calders Formation (Taberner 1983, Hendry *et al.* 1999).

3.2.3 Miliolid micritic arenite

(Terminal Complex, Milany Depositional Sequence)

Lithologies: Miliolid micritic arenite

Occurrence and bed characteristics

The miliolid micritic arenite facies is encountered in a single locality close to the intersection of the N 141 and B 124 roads within Calders village (GR16322692-**Figure 3.1**). This facies occurs stratigraphically below the peloidal wacke/grainstone

facies (**Figure 3.5a, Section 3.2.4**). The lateral extent of this facies cannot be determined due to the lack of laterally continuous exposure. Measured bed thickness is consistently around 1 m. Upper and lower bed contacts are sharp and planar and sedimentary structures are absent (**Figure 3.5a**). *Skolithos* burrows up to 5 cm are common in the field.

Lithological Description

The exposed surface of this facies weathers to a buff/light brown colour (**Figure 3.5a**). Bedding parallel, reddened horizons are visible in outcrop (**Figure 3.5a**). These horizons are up to 35 cm in thickness. Fresh surfaces are a pale blue-grey colour. Siliciclastic grains are monocrystalline unstrained and strained quartz (35 %), with minor lithic grains (10 %) and orthoclase feldspar (10 %). Rare calcite grains are also observed (< 1%). Siliciclastic grains are angular, moderately well sorted with low sphericity. Maximum siliciclastic grainsize is 2 mm (**Figure 3.5b**).

This facies is characterised by a moderate abundance of miliolid foraminifera (<5 %). Miliolids are very low diversity and may be large (up to 1 mm). Miliolids are preserved intact although they are abraded and tests may be reddened (**Figure 3.5b**). Other rare bioclasts associated with the miliolid micritic litharenite facies are small, undifferentiated benthic foraminifera (<1 %) and heavily micritized and abraded silt grade laminar coralline algae fragments (<1%). Internal structure is poorly preserved.

Up to 50 % of this facies consists of dark brown micrite with localised microspar patches (**Figure 3.5b**). Intragranular microspar cements are observed within miliolid tests and iron-rich clay-grade material can be seen to accumulate along mm-scale dissolution seams.

Interpretation: depositional environment

The miliolid micritic arenite facies contains fauna characteristic of a marine environment, although protected conditions, possibly with enhanced salinities, is inferred from the very low faunal diversity and presence of miliolids (Chaproniere 1975, Ghose 1977, Hallock and Glenn 1986, Murray 1991, Geel 2000).

Proximity to a siliciclastic source is inferred from the abundance of siliciclastic material within the miliolid micritic arenite facies. The medium sand

Figure 3.5a Outcrop exposure of the miliolid micritic arenite and peloidal wacke/grainstone facies. Note the slightly reddened horizons, interpreted as possible oxidation surfaces, within the miliolid micritic arenite facies (arrowed). Photograph taken from GR 16322692, facing north. Scale bar=2 m.

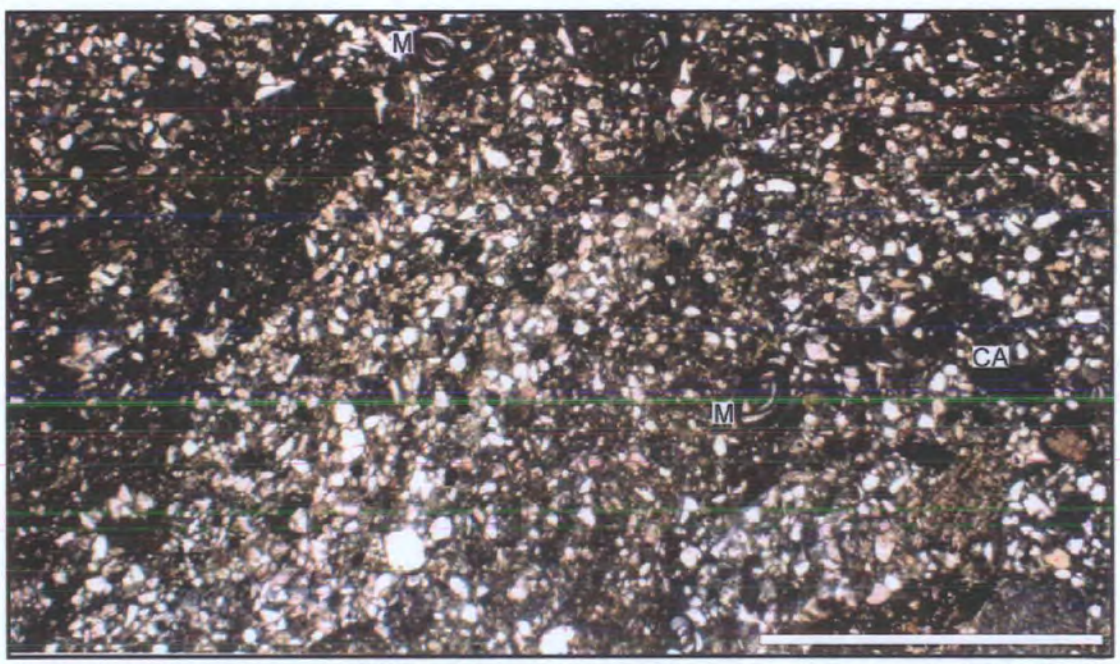


Figure 3.5b Photomicrograph (PPL) of the miliolid micritic arenite facies (sample LCA 78, GR 16322692). **M**=miliolid, **CA**=coralline algae. Light, angular grains are siliciclastic grains. Light and dark patches are interpreted to be a product of bioturbation. Scale bar=2mm.

grade component would have been transported as bedload material. Limited abrasion due to transport or reworking prior to deposition and lithification is inferred from the angular nature of the siliciclastic grains. Low-energy conditions may be deduced by the significant amount of fine-grained carbonate matrix. The micrite is interpreted, at least in part, to be detritus from bioerosion and abrasion of bioclasts.

The bedding-parallel reddened horizons are interpreted to represent temporary subaerial exposure of sediments and oxidation of the muddy matrix. These horizons may be incipient palaeosol horizons, although no distinctive pedogenic structures are observed in thin section and no erosive contacts are observed in outcrop. In developing soil horizons, the patchy transformation of iron mobilized in pore waters to iron oxides in areas with a variable redox potential of groundwater results in mottled textures and red colouration (as well as green, brown and ochre). Red pigmentation may also occur however in relatively humid conditions where the water table is low resulting in oxidising conditions (Besly and Turner 1983). Red pigmentation as a consequence of groundwater evaporation/transpiration is highly unlikely as the climate in this region during the late Middle Eocene was subtropical to temperate (Plaziat 1981, Cavagnetto and Anadon 1996).

In summary, the miliolid micritic litharenite facies represents deposition in a low- to moderate-energy paralic environment undergoing period exposure and oxidation, in proximity to a siliciclastic source. This facies is comparable to the porcellaneous foraminifera micrite arenite facies observed in the Sant Amanc area (**Section 3.6**), although a direct correlation is not possible due to the lack of continuous exposure. This facies is a candidate for a further exposure of the Terminal Complex (see **Section 2.5.3.1**).

3.2.4 Pelletal wacke/grainstone

Lithologies: Pelletal wacke/grainstone

Occurrence and bed characteristics

The pelletal wacke/grainstone facies is encountered in a single locality close to the intersection of the N 141 and B 124 roads within Calders village (GR16322692-**Figure 3.1**). This facies occurs stratigraphically above the miliolid micritic arenite and the main carbonate-siliciclastic succession described in **Section 3.3**. The pelletal wacke/grainstone facies is well bedded with a measured bed thickness of 1 m (**Figure**

3.5a). Observed bed contacts are sharp and planar. Lateral extent of this facies cannot be determined due to lack of exposure.

Lithological description

Exposed surfaces of the pelletal wacke/grainstone facies weather to a buff to grey colour and have a crystalline sugary texture (**Figure 3.6a**). Fresh surfaces are a homogenous blue-grey colour. Pellets are 0.05 to 0.2 mm in diameter and comprise up to 75 % of this facies. Pellets demonstrate ovoid ellipsoidal morphologies with micritic to microcrystalline internal fabrics (**Figures 3.6b** and **c**). Green-brown coloured micritic matrix is patchily distributed, constituting up to 20 % of this facies (**Figure 3.6b**). Intergranular microspar cement is also present (**Figure 3.6**). Iron-rich clays occur concentration long dissolution seams (**Figure 3.6a**). Fossils are very rare in this facies. Small benthic foraminifera up to 0.1 mm comprise less than 1 % of this facies (**Figure 3.6b**).

Interpretation: depositional environment

'Difficult' environmental conditions, with probable enhanced salinity, are inferred from the low faunal diversity of the pelletal wacke/grainstone facies. Modern pellets (and peloids) are a common constituent of Recent shallow subtidal to intertidal environments (Flügel 1972) where they are the product of gastropods or crustaceans. Wilson (1975), Scoffin (1987) and Tucker and Wright (1992) suggest peloids/pellets typify environments with moderate or restricted circulation, although agitated conditions are inferred from the patchy preservation of micrite. Currents were not strong enough however to remove small foraminifera tests.

In summary, the pelletal wacke/grainstone facies was deposited in a restricted shallow water setting under low to moderate energy conditions. Due to its association with the miliolid micritic arenite, it is inferred that this facies is also a component of the Terminal Complex, although no similar lithologies are encountered in the Sant Amanc area (see **Section 3.6**), or described from Terminal Exposure outcrops described from the eastern Vic Basin (Travé 1992, Travé *et al.* 1996).

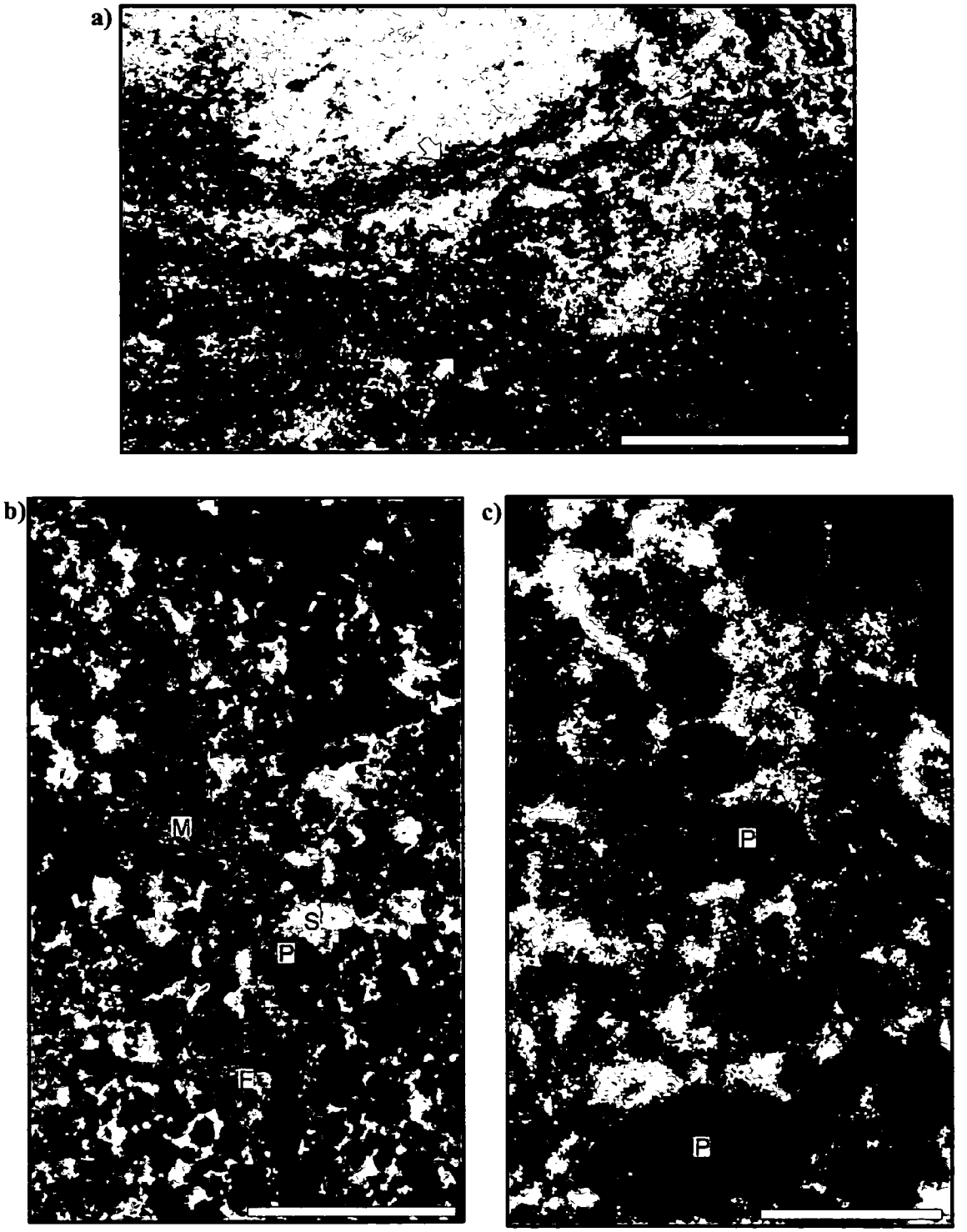


Figure 3.6 Photomicrographs (PPL) of the pelletal wacke/grainstone facies (sample LCA 76, GR 16323692). **a)** Iron-rich, clay-grade material (arrowed) concentrated along dissolution seams. Scale bar=5 mm. b) Variable texture, with micrite (**M**) patchily distributed between pellets (**P**). Intergranular areas have been cemented with calcite spar (**S**). Small, hyaline benthic foraminifera (**F**) are present but very rare. Scale bar =2mm. c) Detailed view of pellets (**P**). The pellets are structurelss in PPL, and are interpreted to be faecal in origin. Scale bar=1mm.

3.2.5 Red mudstone

(Artes Formation, Solsona Depositional Sequence)

Lithologies: Red siltstone
Muddy quartz arenite

Occurrence and bed characteristics

The red mudstone facies is encountered in isolated outcrops along the N 141 road (GR 16052680) and isolated outcrops in the Sant Amanc area (GR 18062680). This facies is stratigraphically the youngest present in the study area, and occurs above the main Calders succession (described in **Section 3.3**) and above the Terminal Complex sediments at Sant Amanc (see **Section 3.6**). It is stressed that the red mudstone facies does not directly overlie the sediments of the main Calders succession. It is inferred from occurrences in the Sant Amanc area and to the far west of the study area, that this facies extends laterally for at least 3 km. This facies is characterised by parallel-laminated red mudrocks with thin (< 10 cm) quartz arenite interbeds (**Figure 3.7a**).

Lithological description

The fresh and weathered surfaces of this facies are a pale red to pink colour (**Figure 3.7a**). Grains are angular, well sorted with low to moderate sphericity (**Figure 3.7b**). Mean grainsize is around 0.1 mm. Siliciclastic grains are predominantly detrital quartz (40 %) with relatively rare lithics (<5 %), mica (<1 %) and calcite (<1 %). Lithic grains are siltstones. Muscovite mica occurs as rare platy grains up to 3 mm, and is often altered to chlorite. Matrix comprises up to 50 % of the red mudstone facies. Matrix consists of iron-rich, clay to fine silt-grade material. Calcite spar intergranular cement is locally developed within samples of this facies. Cement comprises less than 5 % of this facies. The red mudstone facies is unfossiliferous, however, variability of the matrix and disturbance of fine laminae is evident in thin section (**Figure 3.7b**).

Interpretation: depositional environment

Limited abrasion of grains due to transport or reworking prior to deposition and lithification is interpreted from the angular nature of grains. A low energy environment of deposition is inferred by the significant percentage of muddy matrix.

Figure 3.7a Field exposure of the red mudstone facies (assigned to the Artès Formation). Laminated red mudrocks and siltstones occur interbedded with well-lithified quartz arenite. Photograph taken from GR 18062680 (within the Sant Amanc area), facing north. Scale bar=10cm.

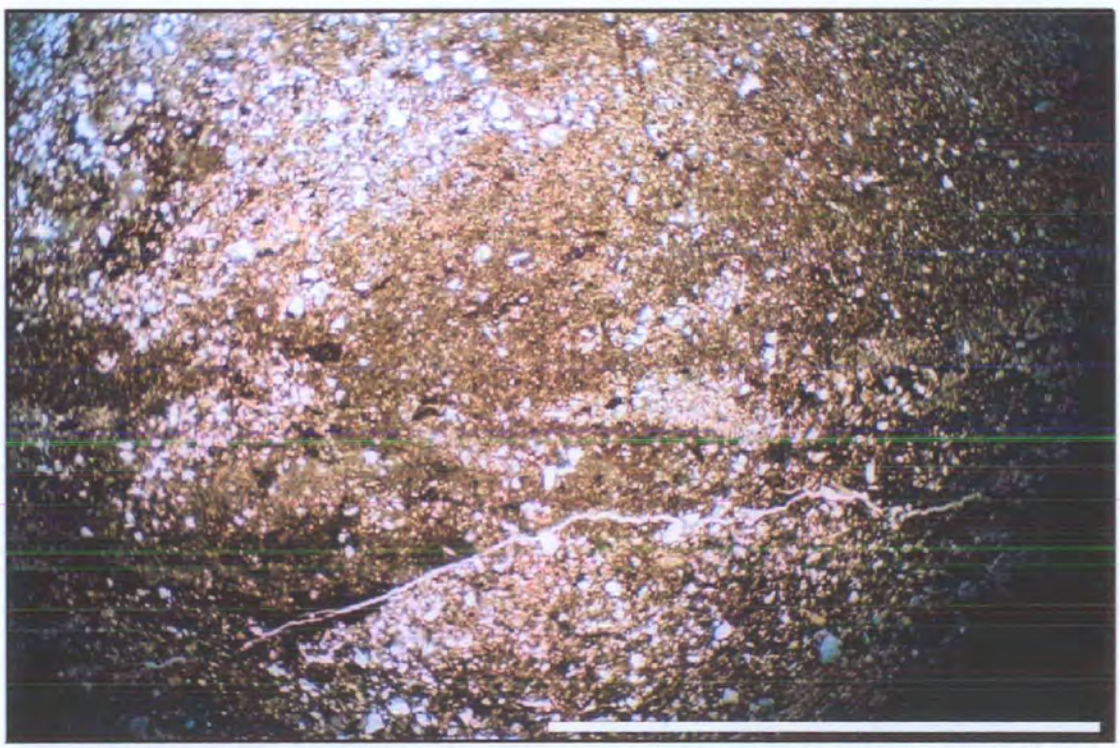


Figure 3.7b Photomicrograph (PPL) of the red mudstone facies (sample LCA 74, GR 16052680). The textural variability of this facies is attributed to pedogenic activity. Scale bar=5mm.

A continental environment of deposition is interpreted by the abundance of iron-rich matrix, the presence of potential pedogenic structures and the absence of marine fauna. Hendry *et al.* (1999) have previously interpreted the red mudstone facies as alluvial sediments. The red mudstone facies is assigned to the Artès Formation.

3.3 Calders facies

The Calders succession is exposed as a series of roughly east-west trending valley systems located along the B 124 road to the south of Calders village (**Figure 3.1**). Individual lithological units may be traced and sampled (where access permits) for several kilometres (**Figure 3.8**).

Three facies groups are defined according to the dominant composition i.e. predominantly carbonate, siliciclastic or mixed carbonate-siliciclastic composition. The facies scheme for all identified facies in the Calders area is presented on **Table 3.2**. The facies scheme is necessarily complicated due to the diversity of components identified in the succession. Limestones are classified according to the schemes devised by Dunham (1962), Embry and Klovan (1971) and Insalaco (1998) (**Appendix 1**). Six carbonate intervals, intercalated with siliciclastics, are identified in the Calders sedimentary succession (**Figure 3.9**).

3.3.1 Carbonate-dominated facies group

Facies included in this grouping are predominantly carbonate in composition, although they may contain a moderate to low siliciclastic component (0 to 34 %). Facies included in this grouping contain a diverse biota dominated by corals, coralline algae and benthic foraminifera. Subordinate biota include echinoids, molluscs, sponges and serpulids. A diversity of carbonate textures identified within this facies group include mixstone, rudstone, floatstone, grainstone and packstone (**Table 3.2**).

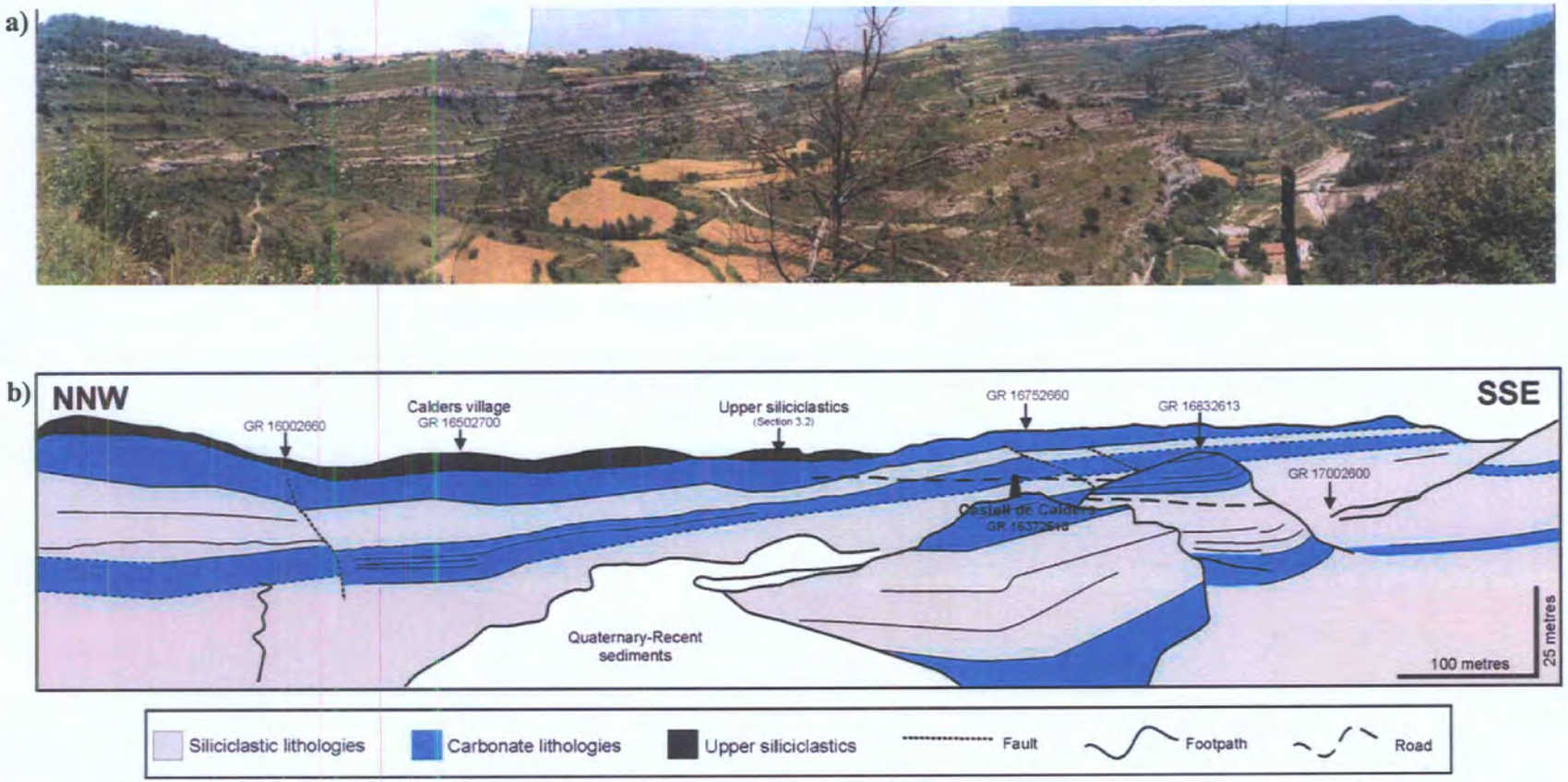


Figure 3.8 Photomontage (a) and interpretation (b) of the studied sedimentary succession at Calders. Although the area has undergone post-depositional faulting, beds are undeformed, and individual units may be traced laterally over 3 km. Photograph taken from GR 16252674, facing approximately NE.

Table 3.2 Calders facies scheme

Facies group and facies	Dominant textures	Bed thickness	Bedding characteristics	Bed contacts	Lateral continuity	Components	Clastic wt. %	Sedimentary structures	Energy level	Environmental interpretation
Carbonate-dominated facies (**Section 3.3.1**)										
Coral mixstone	Mixstone Pillarstone Platestone	Up to 3.5 m	Nodular weathering., clay partings	Gradational lower. Sharp upper	Unknown	Corals, coralline algae, encrusting foraminifera, molluscs.	17 to 27.5	None	Moderate	*In situ* coral growth. Biostrome.
Coral domestone	Domestone	1 to 1.5 m	Nodular weathering, clay partings	Gradational upper, sharp lower	Up to 10 m	Corals, coralline algae, encrusting foraminifera, miliolids, serpulids, texturaliids	11 to 12.5	None	Moderate	*In situ* coral growth. Biostrome.
Coral bioclastic pack/rudstone	Rudstone Packstone Floatstone	Up to 2.65 m	Nodular Crystalline	Sharp, non-erosional upper and lower	Up to 1 km	Corals, coralline alage, encrusting foraminifera molluscs	7.5 to 20.5	Bedding (inferred from alignment of nodules)	Moderate to low	Fore-reef debris
Coral wacke/ floatstone	Wacke/floatstone Floatstone Float/pillarstone Float/platestone	Up to 3 m	Nodular weathering, marl interbeds	Sharp, non-erosional upper and lower	Up to 2 km	Corals, coralline algae, gastropods, bryozoa	12 to 34	Bedding (inferred from alignment of nodules)	Low	Constratal coral growth in a low-energy, muddy setting with significant non-carbonate input
Coralgal foraminifera rudstone	Rudstone	1.2 to 2.4 m	Nodular weathering, clay partings	Sharp, non-erosional upper and lower	Up to 500 m	Corals, coralline algae, encrusting foraminfiera	5 to 21.5	Bedding (inferred from alignment of nodules)	Moderate to high	Proximal fore-reef debris with *in situ* suprastratal coral growth
Coralgal foraminifera float/rudstone	Floatstone Float/rudstone	0.5 to 1.5 m	Nodular weathering, clay partings	Diffuse, gradational upper and lower	Up to 500 m	Corals, coralline algae, encrusting foraminfiera	8 to 12.5	Bedding (inferred from alignment of nodules)	Moderate to low	Distal fore-reef debris with patchy *in situ* coral growth

Table 3.2 (cont.)

Facies group and facies	Dominant textures	Bed thickness	Bedding characteristics	Bed contacts	Lateral continuity	Components	Clastic wt. %	Sedimentary structures	Energy level	Environmental interpretation
Carbonate-dominated facies (**Section 3.3.1**)										
Foralgal pack/ grainstone	Packstone Pack/grainstone Grainstone Grain/rudstone	Up to 2 m	Massive, locally nodular	Sharp but non-erosional upper and lower	> 1 km	Encrusting and articulated coralline algae, diverse benthic foraminifera, *in situ* coral colonies	5.5 to 14.5	None	High to moderate	Shallow water with open marine circulation
Gypsina grainstone	Grainstone Pack/grainstone	1.4 to 1.8 m	Massive, locally nodular	Sharp but non-erosional upper and lower	> 1 km	*Gypsina, Rotorbinella, Calcarina, Amphistegina,* coralline algae	7.5 to 21.5	None	High to moderate	Shallow water with open marine circulation, vegetated shoals
Miliolid coralline algae pack/ grainstone	Packstone Pack/grainstone	Up to 1 m	Nodular weathering, crystalline texture	Sharp lower, gradational upper	Up to 500 m	Miliolids, *Orbitolites,* coralline algae	7 to 26.5	Bedding defined by alignment of nodules	Moderate	Shallow water with partially restricted marine circulation
Mixed carbonate-siliciclastic larger benthic foraminifera facies (**Section 3.3.2**)										
Nummulites siliciclastic pack/grainstone	Packstone Grainstone Pack/grainstone Pack/wackestone	1 to 3.4 m	Tabular beds. Sandy texture	Upper and lower bed contacts with carbonates may be dissolutional	> 2 km	*Nummulites, Gypsina, Discocyclina,* coralline algae	14 to 50	Planar cross-stratification	Moderate to high, quiet periods	Reworked *Nummulites* accumulation
Nummulites Discocyclina siliciclastic packstone	Packstone Wacke/packstone	Up to 1.7 m	Tabular beds. Sandy texture	Gradational upper and lower	> 500 m	*Discocyclina, Nummulites, Operculina*	14 to 38.5	Bedding, chaotic stacking of foraminifera tests	Low	*Discocyclina* accumulation reworked by low-energy currents

Table 3.2 (cont.)

Facies group and facies	Dominant textures	Bed thickness	Bedding characteristics	Bed contacts	Lateral continuity	Components	Clastic wt. %	Sedimentary structures	Energy level	Environmental interpretation
Mixed carbonate-siliciclastic larger benthic foraminifera facies (**Section 3.3.2**)										
Operculina mudstone	Mudstone Wackestone	Up to 1.6 m	Weathers recessively	Sharp, non-erosional	1 km	*Operculina*, delicate bivalves, undifferentiated fine-grained fossil debris	35 to 37.5	None	Very low	Low-energy open marine within the lower portion of the photic zone
Nummulites coralline algae siliciclastic packstone	Packstone Wacke/packstone Pack/grainstone	0.85 to 1 m	Sandy texture	Gradational lower, sharp non-erosive upper	> 1 km	*Nummulites*, coralline algae, foralgal rhodoliths, siliciclastic grains	15 to 26.5	None	Moderate to high	Moderate energy, open marine setting with high clastic input and unstable substrates
Coralgal foraminifera siliciclastic float/rudstone	Float/rudstone	1.5 m	Massive to locally nodular beds	Sharp, non-erosional upper and lower	> 500 m	Corals, coralline algae, encrusting foraminifera, siliciclastic grains	22 to 23.5	Bedding (defined by alignment of nodules)	Moderate to low	Patchy coral development within an open marine, agitated setting with a siliciclastic input
Bioclastic siliciclastic-dominated facies (**Section 3.3.3**)										
Gypsina calcarenite	Sandstone Pack/wackestone	1 to 2.5 m	Sandy texture, bioturbation	Sharp, non-erosional upper and lower	> 1 km	*Gypsina*, *Rotorbinella*, echinoids, bryozoa	35.5 to 69	M-scale cross-stratification	Moderate to high	Shallow marine, high-energy, vegetated siliciclastic shoal
Nummulites arenite	Sandstone	1.15 to 1.65	Sandy texture, bioturbation	Gradational upper and lower	500 m	Siliciclastic grains, *Nummulites*	Up to 55	Bedding	Moderate	Shallow marine agitated siliciclastic shelf with sparse biota

Table 3.2 (cont.)

Facies group and facies	Dominant textures	Bed thickness	Bedding characteristics	Bed contacts	Lateral continuity	Components	Clastic wt. %	Sedimentary structures	Energy level	Environmental interpretation
Bioclastic siliciclastic-dominated facies (**Section 3.3.3**)										
Bioclastic siltstone	Siltstone Wackestone	0.4 m	Weathers recessively, locally nodular	Upper and lower sharp and non-erosional	Unknown	Silt-grade siliciclastic grains, *Nummulites*, coralline algae, echinoids	Up to 45	Massive	Very low	Very low-energy, open-marine setting with a minor bioclastic and siliciclastic input from high-energy parts of the shelf
Siliciclastic-dominated facies (**Section 3.3.4**)										
Cross-stratified calcarenite	Sandstone Siltstone	0.1 to 3 m	Sigmoidal and tabular bedding	Upper and lower sharp, often erosional.	> 4 km	Quartz, lithics, feldspar, calcite, dolomite, micritic matrix. Rare bioclasts include echinoids, gastropods, coralline algae, small hyaline benthic foraminifera	Up to 75	Metre-scale cross-stratification, swaley cross-stratification, symmetrical ripple lamination	Low to high	Prograding shallow marine siliciclastic shelf
Lenticular polymict conglomerate	Conglomerate Conglomeratic sandstone	Up to 1.4 m	Lenticular. Up to 20 m wide.	Lower contact erosional. Upper contact gradational	> 100 m	Poorly sorted lithic pebbles supported by a coarse sand-grade calcarenite matrix	Up to 100	Lateral accretion surfaces	High	Conglomerate-infilled channels

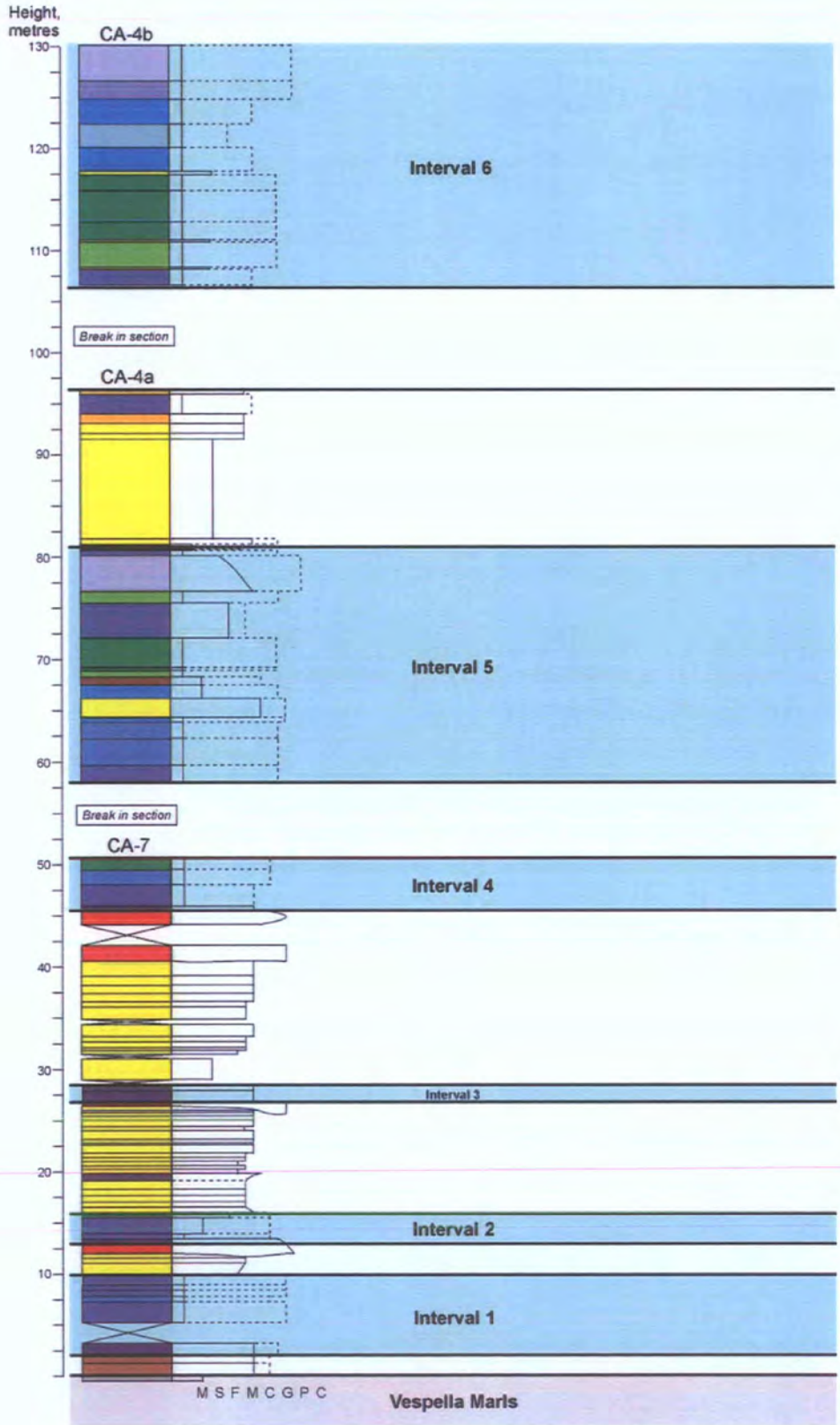


Figure 3.9 Main carbonate intervals referred to in the text (highlighted in pale blue) situated within the predominantly siliciclastic sedimentary succession at Calders. Logs CA-7, CA-4a and CA-7b illustrate the succession exposed along the B124 road (**Figure 3.1**)

3.3.1.1 Coral mixstone

Lithologies: Coral mixstone
Coral platestone

Occurrence and bed characteristics

Stratigraphically, the coral mixstone facies is identified in the upper portion of the studied succession at Calders (within carbonate intervals 5 and 6-**Figure 3.9**). This facies is exposed along the B 124 road between GR 16402692 and 16502675 (log CA-4b, bed 13), and GR 16752642 (log CA-4a, bed 13) (**Figure 3.1**). The lateral extent of this facies cannot be determined through the absence of laterally continuous exposure. Measured bed thickness ranges from 3.25 to 3.5 m. Lower bedding contacts with the coral domestone facies (log CA-4b, beds 12 and 13) are gradational over a vertical distance of 50 cm. Upper contacts are not exposed in the study area.

The coral mixstone facies has a nodular weathered texture. Each nodule typically corresponds to a coral colony with blue-green to brown mud surrounding each nodule. This facies has a brecciated appearance towards the top of the exposure due to post-depositional faulting.

Lithological description

Exposed surfaces of the coral mixstone facies weather to a light brown-grey colour (**Figure 3.4**). Fresh surfaces are a pale blue-grey colour and with a crystalline texture. This facies is characterised by an abundance of *in situ* coral colonies. The total coral skeletal volume of this facies ranges from 50 to 55 %. Coral morphologies observed are branching, head, platy, foliaceous and massive (**Figure 3.10**). Coral colony size is highly variable (from 3 cm to 0.75 m). Platy coral colonies may become increasingly abundant towards the top of this facies (log CA-4b, bed 13). Coral genera identified are *Porites*, *Goniopora*, *Acropora, Siderastrea, Astreopora, Alveopora, Stylophora* and *Favites* (cf. Vilaplana 1977). A pack/rudstone matrix containing abundant angular coral debris is observed in between the *in situ* colonies. Beds of this facies contain up to 30 % pack/rudstone matrix. Coral fragments up to 5 cm are poorly sorted.

Branching coral colonies are encrusted by coralline algae (**Figures 3.11a** and **b**) and may be extensively bored by *Lithophaga* (**Figures 3.11 c** and **d**). Borings are cylindrical in cross section and almond shape in longitudinal section. *Lithophaga* tests

may be preserved within coral colonies (**Figure 3.11c**). Elliptical rhodoliths up to 5 cm in diameter are present in this facies (**Figure 3.12**). Bioclasts act as nuclei for growth of tight, concentric algal laminae. Thin branches, up to 1 cm in length and 2-3 mm thick, often succeed this initial concentric layer (**Figure 3.12**). A final tight, concentric layer is the latest growth phase. Large encrusting foraminifera are present within coral debris. Foraminfera present are *Haddonia* (**Figure 3.11b**), victoriellids (**Figure 3.11e**), *Fabiania* and acervulinids. *Fabiania* and *Haddonia* tests are up to 4 mm and are preserved intact.

Dark-brown muddy matrix comprises up to 50 % of this facies, and is preserved locally in pockets between coral colonies. Seventeen to 27.5 wt. % of the matrix is composed of clay-grade non-carbonate material. Relatively rare fine sand-grade siliciclastic grains (<2 %) are monocrystalline and polycrystalline quartz, orthoclase feldspar (1 %) and lithic grains (1 %). Lithic grains are siltstone and quartz mica schist. Feldspars commonly have a brown, altered appearance in thin section. Siliciclastic grains are angular, poorly sorted and have a low sphericity.

Other fossils present in the matrix of the coral mixstone facies include small hyaline benthic (0.5 %) and texturaliid foraminifera (0.5 %), green calcareous algae (0.5 %-**Figure 3.11d**) and molluscan debris (2 %). Rare specimens of *Amphistegina* are also observed. Well-preserved, robust *Amphistegina* tests are up to 1 mm in diameter.

Diagenesis

Originally aragonitic bioclasts, such as corals and green algae, have been leached with the resultant mouldic pores completely occluded with inclusion-free, drusy calcite spar cement (**Figure 3.11b, d** and **e**). Drusy calcite cements have also occluded macropores within *Haddonia* tests (**Figure 3.11b**). It was not possible to distinguish intergranular cements from the muddy matrix using standard petrological techniques employed in this study, although some degree of cementation and/or matrix recrystallisation is inferred from the nodular appearance of this facies in outcrop. Compaction of sediments is evident from dissolutional contacts between nodules where clays are concentrated. This is evident in thin section with iron-rich clays concentrated along dissolution seams between coral fragments (**Figure 3.11d**).

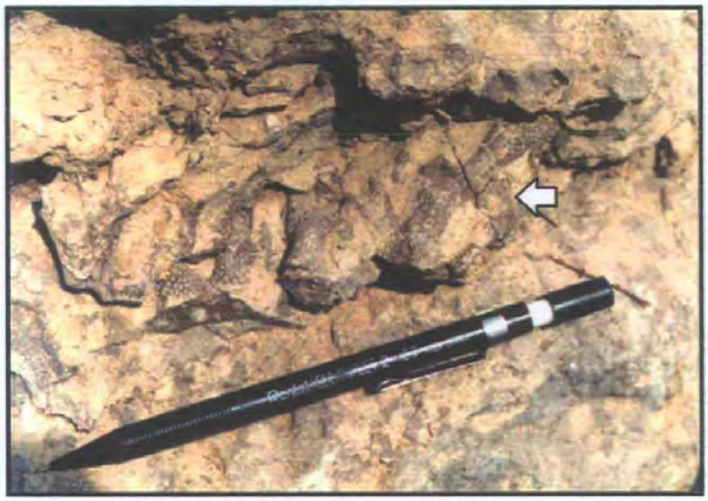


Figure 3.10a Small branching *Acropora* colony (arrowed) within the coral mixstone facies (log CA-4b, bed 13). The colony is preserved intact, and is situated within a bioclastic wackestone matrix. Pencil=14cm.

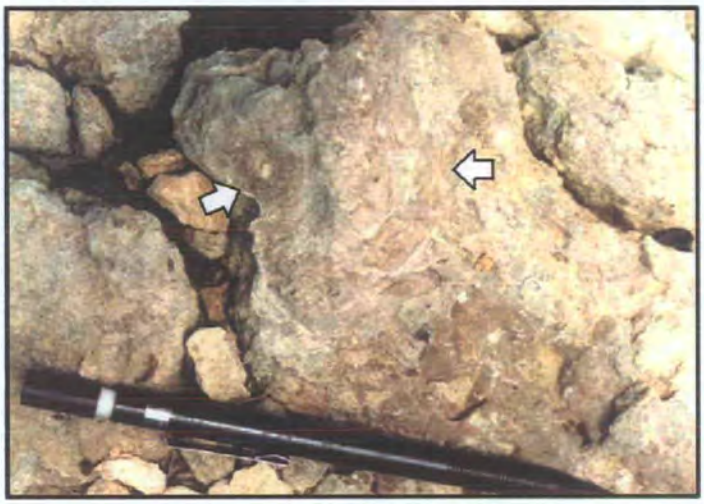


Figure 3.10b Foliaceous *Porites* colony (arrowed) within the coral mixstone facies (log CA-4a, bed 13). Pencil=14cm.

Figure 3.10c Large head coral colony (*Sinuosiphyllia*) identified at the transition between the coral domestone and coral mixstone facies (log CA-4b, bed 13). Pencil=14cm.

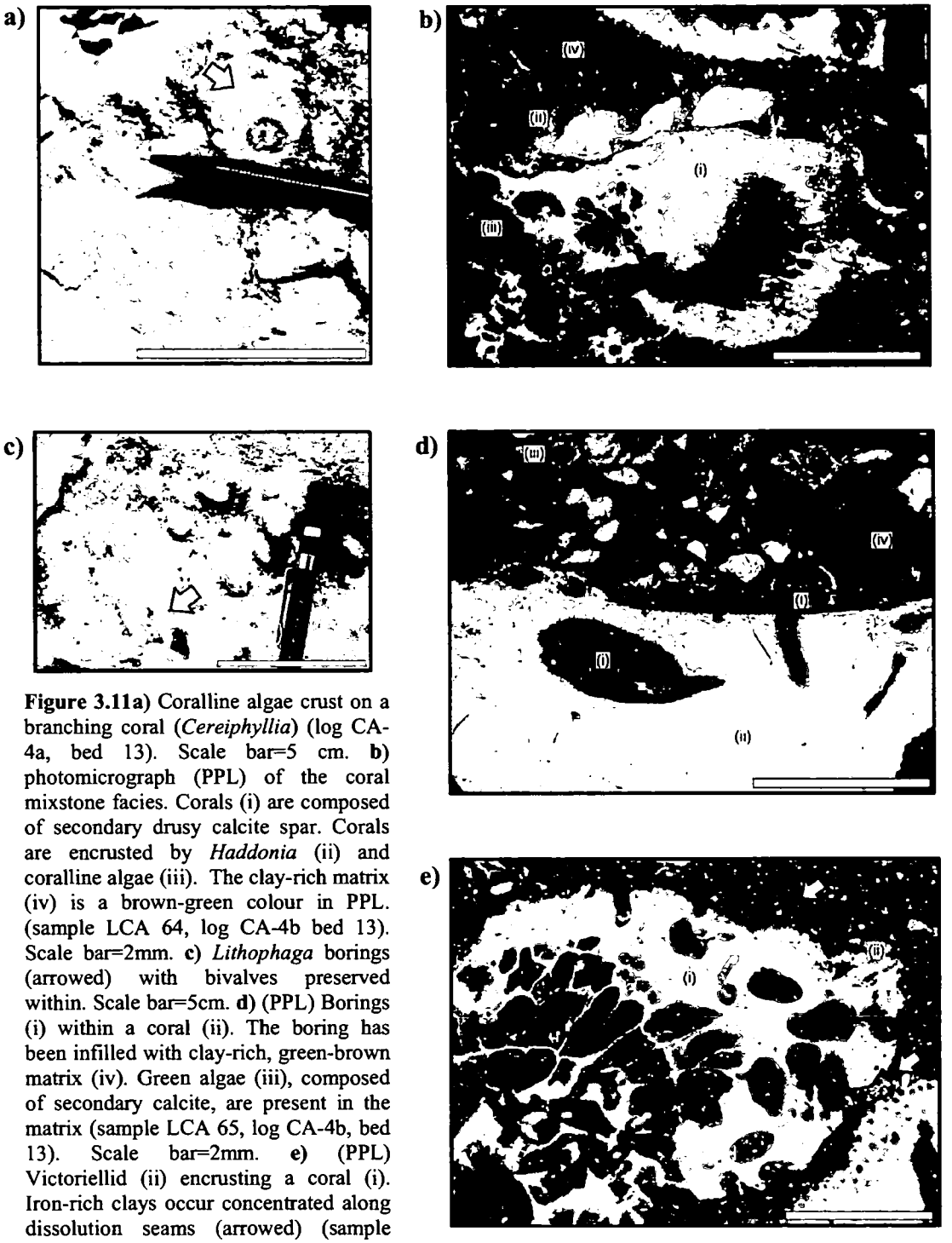


Figure 3.11a) Coralline algae crust on a branching coral (*Cereiphyllia*) (log CA-4a, bed 13). Scale bar=5 cm. **b)** photomicrograph (PPL) of the coral mixstone facies. Corals (i) are composed of secondary drusy calcite spar. Corals are encrusted by *Haddonia* (ii) and coralline algae (iii). The clay-rich matrix (iv) is a brown-green colour in PPL. (sample LCA 64, log CA-4b bed 13). Scale bar=2mm. **c)** *Lithophaga* borings (arrowed) with bivalves preserved within. Scale bar=5cm. **d)** (PPL) Borings (i) within a coral (ii). The boring has been infilled with clay-rich, green-brown matrix (iv). Green algae (iii), composed of secondary calcite, are present in the matrix (sample LCA 65, log CA-4b, bed 13). Scale bar=2mm. **e)** (PPL) Victoriellid (ii) encrusting a coral (i). Iron-rich clays occur concentrated along dissolution seams (arrowed) (sample LCA 65, log CA4b, bed 13)

Interpretation: Depositional environment

The coral mixstone facies contains a marine faunal assemblage. Open marine conditions are inferred from the diversity of biota and the presence of *in situ* corals.

Deposition within the photic zone is concluded from the abundance of *in situ* coral colonies. Although tightly packed, no discernable topography developed. Large encrusting foraminifera (*Haddonia, Fabiania* and victoriellids) colonized cryptic environments within coral rubble (cf. Romero 2001, Romero *et al.* 2002).

Variable depositional energies are inferred from the changing growth forms of rhodoliths. The switch from tight, concentric algal laminae to delicate branching growth is consistent with a switch from relatively high to low energy conditions (cf. Bosence 1983). It is postulated that fine-grained siliciclastic input accompanied the higher-energy periods, and settled from suspension as water energy decreased. The development of thin, laterally extensive coralline algae crusts on corals is consistent with a lack of competition for space either due to enhanced nutrients or low incident light (Minnery *et al.* 1985, Minnery 1990). The high incidence of bioerosion in the coral mixstone facies may be an indicator for slightly enhanced nutrient levels (Hallock 1988), although the abundance of corals indicates oligotrophic conditions were the norm.

A progressive decrease in incident light in the final carbonate interval, related to either increasing water depths or enhanced turbidity, is interpreted from the switch from a diversity of coral morphologies at the base of this facies to the abundance of platy coral morphologies at the top. The development of flat, platy morphologies is a response of a photosymbiont-bearing organism increasing the surface area dedicated to harvesting light for energy (Titlyanov and Latypov 1991).

In summary, the coral mixstone facies is interpreted as a biostrome influenced by variable depositional energies and fine-grained siliciclastic input. Faunal assemblages dominated by corals, coralline algae and large encrusting foraminifera, comparable with the coral mixstone facies, are described from the Eocene of the Igualada Basin (Romero 2001, Romero *et al.* 2002). These assemblages have been interpreted as a barrier reef complex, and lack of significant topographic relief is attributable to lack of wave exposure (Romero *et al.* 2002).

3.3.1.2 Coral domestone

Lithologies: Coral domestone

Occurrence and bed characterisitics

The coral domestone facies is exposed at the top of the Calders section only (carbonate interval 6-**Figure 3.9**) and can be traced laterally for 10 m. Measured thickness ranges from 1 to 1.5 m. The lower contact of this facies with the foralgal grainstone is sharp but non-erosional. The upper contact with the coral mixstone is gradational, with a gradual diversification in coral colony morphology over a distance of approximately 50 cm.

This facies has a nodular weathered appearance in outcrop. Nodules are up to 75 cm and often correspond to an individual coral colony. Green-grey coloured clays occur between nodules.

Lithological description

The exposed surface of this facies weathers to a grey-brown colour. Fresh surfaces are a blue-green colour. The coral domestone facies is characterised by an abundance of small domal coral colonies (**Figures 3.13a** and **b**). Colonies are closely packed apparently forming a framework. The total coral skeletal volume ranges from 50 to 55 %. Coral genera identified include *Stylophora, Actinacis* and undifferentiated faviids. Coral colony size ranges from 15 to 50 cm and there is a noticeable vertical increase in size. Lithophagid bivalve borings are evident on coral colonies (**Figure 3.13b**).

Beds of the coral domestone facies are composed of up to 30 % bioclastic pack/rudstone rich in coral fragments. The pack/rudstone matrix occurs in pockets between *in situ* coral colonies. The matrix contains abundant dark brown mixed carbonate siliciclastic mud (11.5 to 53 %) (**Figure 3.14a** to **c**). Up to 25 wt.% of the mud is non-carbonate in composition. The matrix contains coral fragments (20 to 50.5 %), molluscs (2.5%), coralline algae (2 to 11.5%), miliolids (0 to 5%), texturaliids (0 to 1.5%), bryozoa (<1%) and echinoids (0.5 to 3%). All fossils in the matrix are fragmented. Coral fragments are encrusted by coralline algae, larger foraminifera (victoriellids) and serpulids (**Figures 3.14a** to **c**).

Diagenesis

Originally aragonitic coral skeletons have been leached, with the resultant mouldic pores completely occluded with coarse, inclusion-free, drusy calcite spar cement (**Figure 3.14a**). Fine-grained intragranular calcite cements have also occluded porosity within foraminifera tests (**Figure 3.14a**). Locally, the muddy matrix has been recrystallised into inclusion-rich microspar (**Figure 3.14c**). The concentrations of green-grey clays around nodules are interpreted as macro-dissolution seams.

Interpretation: depositional environment

The coral domestone facies contains a marine faunal assemblage. Normal marine conditions within the photic zone are inferred from the abundance of *in situ* corals and the general diversity of stenohaline biota. The corals form a low-density framework although there is no evidence of topographic relief. Victoriellid encrusting foraminifera inhabited cryptic environments within coral debris (Romero 2001, Romero *et al.* 2002).

Shallow water depths with high incident light are concluded from the dominance of domal coral morphologies (Reiss and Hottinger 1984). Low-energy conditions are inferred from the abundance of fine grained matrix and the presence of texturaliid foraminifera that have a tendency to undergo post-mortem disaggregation in high-energy conditions (Goureau and Goureau 1973). Fine-grained, non-carbonate material was deposited from suspension. Enhanced nutrient levels, possibly related to this non-carbonate input, are inferred from the high incidence of bioerosion in this facies (Hallock 1988).

In summary, the coral domestone facies is interpreted as a low-energy, shallow marine coral biostrome with a moderate fine-grained non-carbonate input.

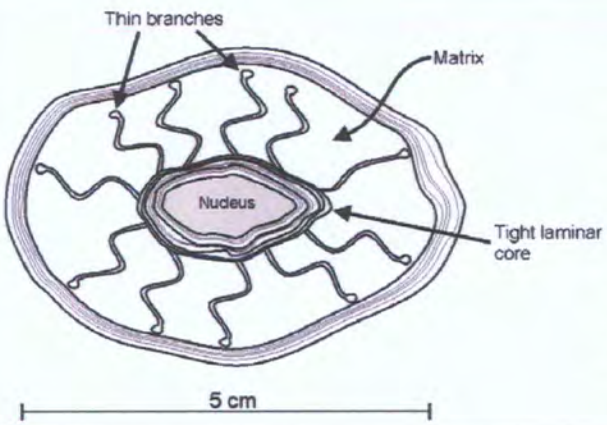


Figure 3.12 Sketch of a typical elliptical rhodolith found within the coral mixstone facies. The nucleus is often a coral fragment. The initial growth is tight concentric laminar, followed by delicate branches. A switch back to tight, concentric laminar growth represents the final growth stage. These variations in growth texture are interpreted to represent relative changes in water energy and possibly non-carbonate sediment input.

Figure 3.13a An *in situ* faviid coral colony demonstrating domal growth morphology, characteristic of the coral domestone facies. Log CA-4b, bed 12. Scale bar = 5cm.

Figure 3.13b Lithophagid borings (arrowed) within an *in situ* domal coral colony. Clays (stained yellow in this photograph) surround nodules in this facies. Log CA-4b, bed 12. Scale bar = 5cm

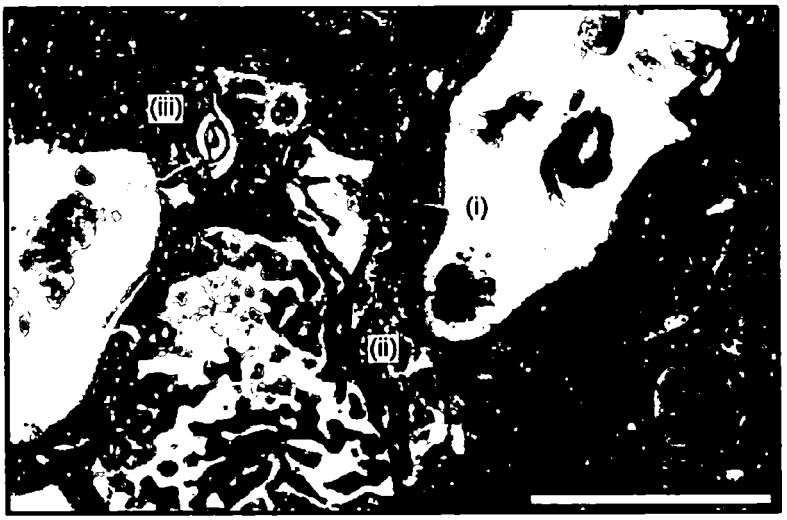


Figure 3.14a. Photomicrograph (PPL) of the coral dome- stone facies. Corals (*Stylophora*-i) are composed of secondary drusy calcite, and may be encrusted by coralline algae (ii). Miliolids (iii) are present in the matrix. Porosity within miliolid chambers has been occluded by drusy calcite cement (sample LCA 60, log CA-4b, bed 13). Scale bar=2mm.

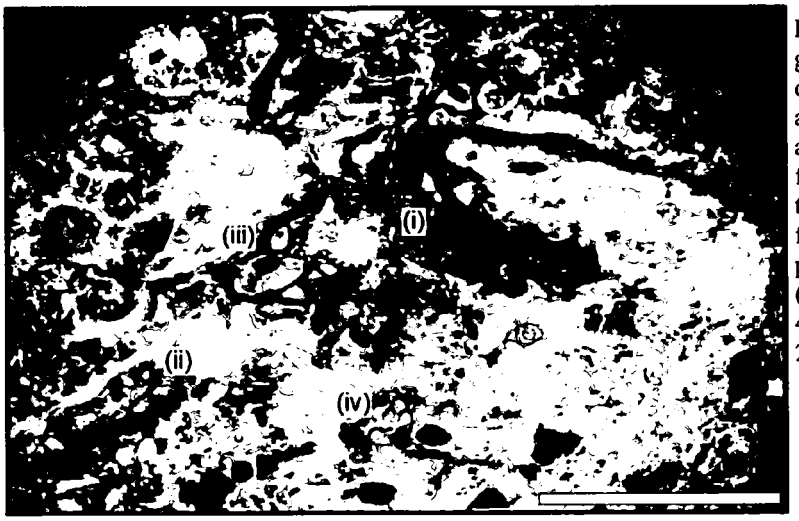


Figure 3.14b. Photomicrograph (PPL) of the coral dome- stone facies. Corals are encrusted by coralline algae (i) and victoriellid foraminifera (ii). Worm tubes (iii) and texturaliid foraminifera (iv) are present in the matrix (sample LCA 61, log CA-4b, bed 13). Scale bar= 2mm

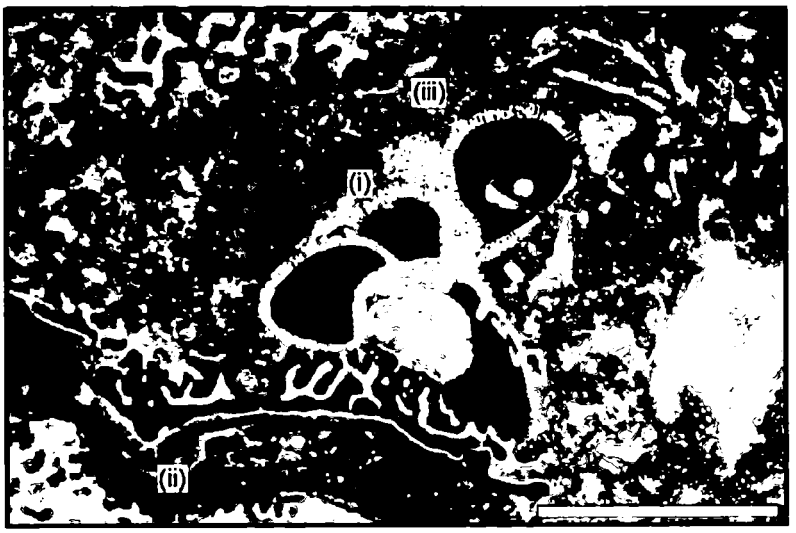


Figure 3.14c. Detailed view (PPL) of the coral dome-stone facies illustrating a large victoriellid foram (i) and a *Lithoporella* crust (ii) on a coral fragment (*Actinacis*). The matrix is locally composed of neomorphic calcite spar (iii) (sample LCA 63, log CA-4b, bed 13). Scale bar= 0.5mm

3.3.1.3 Coral bioclastic pack/rudstone

Lithologies: Coral bioclastic rudstone
Coral bioclastic pack/rudstone
Coral bioclastic wacke/packstone

Occurrence and bed characteristics

The coral bioclastic pack/rudstone facies is encountered in all of the carbonate intervals illustrated on **Figure 3.9**. Measured bed thickness is up to 2.65 m, and beds may be traced laterally (along strike) for several hundred metres. This facies has a nodular weathered appearance. Nodules are up to 10 cm in diameter and are aligned. Alignment of nodules is inferred to be remnant bedding, with beds dipping up 5° consistently towards the northwest. Upper and lower bedding contacts are sharp and non-erosional.

Lithological description

The exposed surface of the coral bioclastic pack/rudstone facies weathers to a buff to grey colour. Fresh surfaces are a pale blue-grey colour. This facies is characterised by sparsely distributed colonial corals situated within a bioclastic pack/rudstone. The estimated coral skeletal volume of this facies is 25 to 30 % and corals are not forming a framework. Coral colonies up to 35 cm occur *in situ*, overturned (**Figure 3.15a**) and fragmented, and are often encrusted by coralline algae, larger foraminifera (i.e. *Gypsina, Haddonia* and victoriellids) and bryozoa.

The bioclastic pack/rudstone matrix is dominated by coral fragments (2.5 to 34.5 %) that range from 2 mm to 15 cm in length (**Figures 3.15b** and **c**). Coral fragments often form the nuclei for rhodoliths. Rhodoliths are up to 7.5 cm in diameter, and show the same growth characteristics identified in the coral mixstone and coral domestone facies (**Figure 3.12**) with an initial concentric laminar crust succeeded by delicate branches. Elliptical rhodoliths up to 6 cm in diameter with thick columnar branches are also present in this facies.

The coral bioclastic pack/rudstone facies contains a very diverse foraminifera assemblage including *Amphistegina* (0 to 3 %), *Gypsina* (0 to 7.5 %), miliolids (0.5 to 2 %), texturaliids (0 to 1.5 %) and large encrusting foraminifera including victoriellids, *Haddonia*, *Fabiania* and *Chapmanina* (0 to 4.5 %). Small, robust forms of *Amphistegina* up to 1.2 mm in diameter have relatively thick test walls

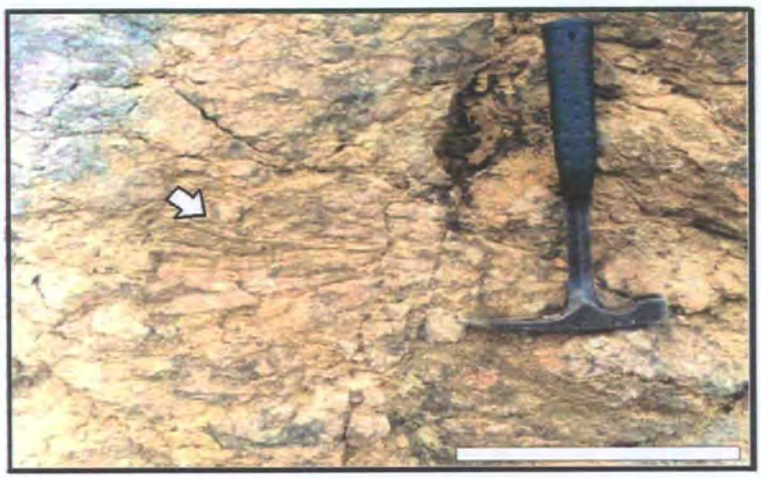


Figure 3.15a Large, overturned branching coral colony within the coral bioclastic pack/rudstone facies (log CA-4a, bed 2) Scale bar=50cm

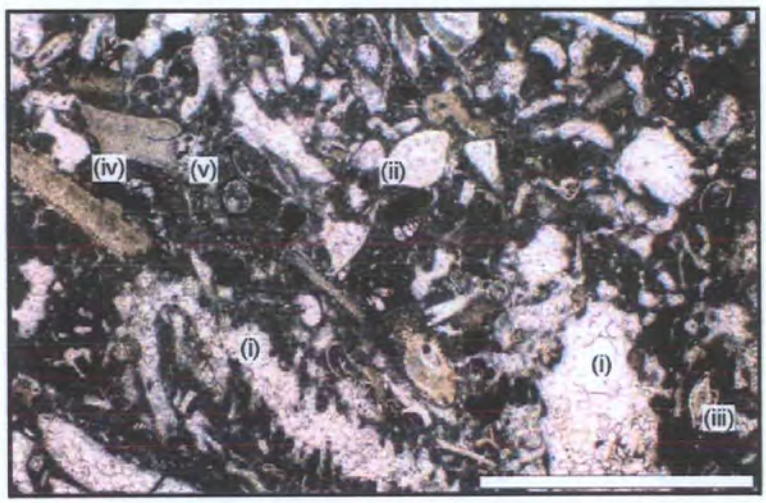


Figure 3.15b. Photomicrograph (PPL) of the coral bioclastic pack/rudstone facies. This facies contains abundant coral fragments (i), with small, robust *Amphistegina* tests (ii), small *Nummulites* (iii), echinoids (iv) and worm tubes (v) (sample LCA 99c, log CA-10, bed 1). Scale bar=2mm

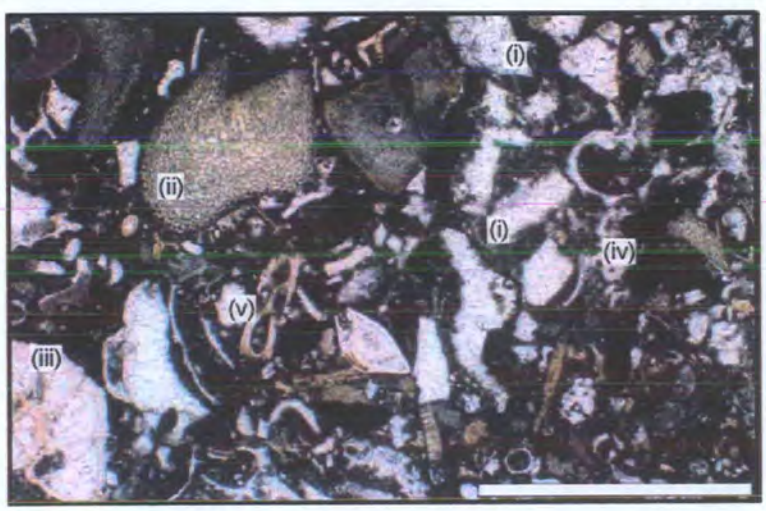


Figure 3.15c Photomicrograph (PPL) of the coral bioclastic pack/rudstone facies. Small, abraded coral fragments (i) are composed of secondary drusy calcite spar. Larger encrusting foraminifera present include *Gypsina* (ii) and victoriellids (iii). Abraded coralline algae (iv) and bryozoa (v) are also abundant (sample LCA 38, log CA-4a, bed 2). Scale bar= 2mm

a)

b)

c)

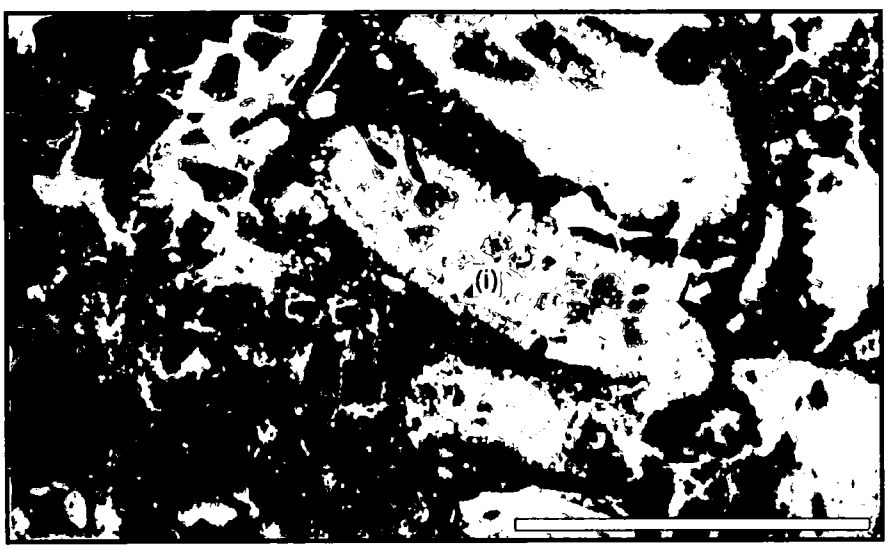


Figure 3.16 a) Small solitary coral (arrowed) within the coral bioclastic pack/rudstone facies (log CA-4a, bed 2). Scale bar=5cm

b) Detailed photomicrograph (PPL) of a foraminifera crust on a coral fragment composed of *Haddonia* (i) and a victoriellid (ii). Note the drusy calcite spar cement within foraminifera chambers (arrowed) (sample LCA 19, log CA-4a, bed 2). Scale bar=1 mm

c) Photomicrograph (PPL) of large coral fragments and bryozoa (i) floating in a dark, muddy matrix. Small microfractures (arrowed) are cemented with drusy calcite (sample LCA 205, log CA-7 bed 5). Scale bar=2mm

(**Figure 3.15b**). *Amphistegina* tests occur intact and as angular fragments in the matrix. *Gypsina* tests, with rounded and laminar morphologies, up to 2 mm in diameter, occur intact and as angular to highly abraded fragments (**Figure 3.14c**). Victoriellid foraminifera are preserved intact, with tests up to 2.5 mm in diameter (**Figures 3.15c and 3.16b**). Solitary corals (**Figure 3.16a**) are also present in this facies but are considered an insignificant component of this facies.

Other bioclasts represented in this facies include detached, laminar crusts of coralline algae (0 to 12.5 %), echinoids (1 to 5 %), molluscs (0 to 4.5 %), sponge spicules (<0.5 %) and fenestrate bryozoa (**Figure 3.16c**). In general, bioclasts are sub-angular and poorly sorted. Peloids, interpreted as highly abraded coralline algae fragments, are particularly common in some samples of this facies (up to 19 %).

A combination of micrite and non-carbonate mud comprises 25.5 to 65 % of the coral bioclastic pack/rudstone facies. The total non-carbonate content ranges from 7.5 to 20.5 wt. %. Minor silt-grade quartz grains (<1 %) are angular and very well sorted.

Diagenesis

Post-depositional cementation of this facies is inferred from the nodular appearance in outcrop, although intergranular cements cannot be differentiated from the muddy matrix. Green-grey clays locally preserved between nodules are interpreted as macro-dissolution seams where non-carbonate material has become concentrated through compaction and dissolution. Originally aragonitic organisms have been leached with the resultant mouldic pores completely occluded with clear drusy calcite spar cement (**Figures 3.15c** and **d**). Drusy calcite cements are also observed within foraminifera tests (**Figure 3.16b**).

Interpretation: depositional environment

The coral bioclastic pack/rudstone facies contains a marine faunal assemblage. Normal marine conditions are inferred from the abundance of stenohaline biota (Murray 1991). Deposition within the photic zone is inferred from the presence of symbiont-bearing foraminifera (*Amphistegina*) and *in situ* coral colonies (Hallock and Glenn 1985 and Leutenegger 1984). Coral colonies have a low coral skeletal volume in this facies and do not form a framework.

Low depositional energies are inferred from the abundance of fine-grained matrix and clay to silt-grade siliciclastic material that was deposited from suspension. However, agitation of this environment is indicated from the presence of rhodoliths (Bosence 1983), with energies high enough to fragment and disperse skeletal organisms. The robust, lenticular morphology of *Amphistegina* is also indicative of high-energy/agitated conditions within the shallow to intermediate parts of its depth range (Hallock and Glenn 1985 and Leutenegger 1984). It is concluded that a significant proportion of bioclasts are derived from a relatively high-energy reef setting, probably in relative shallow water, and were transported into this fore-reef setting during storm events and/or through the effects of gravity. *Amphistegina,* moderately abundant in this facies, is commonly recorded in Recent fore-reef settings (Hallock and Glenn 1986, Li and Jones 1997, Li *et al.* 1997). Large, encrusting foraminifera (*Haddonia, Chapmanina, Fabiania*) lived within cryptic environments within the coral debris (Franqués-Faixa 1996, Romero 2001, Romero *et al.* 2002).

In summary, the coral bioclastic pack/rudstone facies is interpreted as debris derived from a nearby coral reef environment, deposited in a fore-reef setting. This facies has similarities with SMF 6 of Wilson 1975, however the coral bioclastic pack/rudstone facies contains a significant percentage of fine-grained matrix and therefore may be the lower energy/distal equivalent.

3.3.1.4 Coral wacke/floatstone

Lithologies: Coral floatstone
Coral wacke/floatstone
Coral float/platestone
Coral float/pillarstone

Occurrence and bed characteristics

The coral wacke/floatstone facies is common, and is encountered in all of the carbonate intervals illustrated on **figure 3.9**. This facies characteristically succeeds a mixed carbonate-siliciclastic facies (described in **Section 3.3.2**).

This facies is characterised by poorly exposed units with a nodular weathered appearance (**Figure 3.17a**). Measured bed thickness is up to 3 m (although the contacts of this facies are poorly exposed), and beds may be traced laterally for up to 1 km. Visible upper and lower bedding contacts are sharp but non-erosional. Nodules

are 5 to 15 in length and are often aligned (**Figure 3.17a**). Nodules are set within blue-grey marl matrix that weathers to a pale grey/white colour.

Lithological description

The exposed surface of the coral wacke/floatstone facies has a finely crystalline texture and weathers to a white to pale grey colour. Fresh surfaces are a pale green-grey to blue-grey colour.

This facies is characterised by the occurrence of sparse *in situ* delicate coral colonies up to 15 cm in diameter (**Figure 3.17b** and **c**). Coral skeletal volume is up to 10 %, and colonies do not make up a framework. Branching coral morphologies predominate, although platy forms are identified in carbonate interval 2 (**Figure 3.9**). In the western limit of the study area, the first carbonate interval is dominated by laminar and branching forms of *Cyathoseris* (log CA-8, bed 2). Coral fragments, comprising up to 20 % of the matrix, are up to 10 cm in length but more typically are around 2 mm. *In situ* corals and coral fragments may demonstrate thin (<1 mm), laterally extensive, laminar coralline algae crusts (**Figure 3.17c**). Crusts may also comprise foraminifera (*Gypsina*), and bryozoa. Laminar fragments of coralline algae are present in the matrix (0 to 2.5 %). Coralline algae fragments show good preservation of internal structure.

Other bioclasts represented in this facies are *Amphistegina* (< 0.5 %), miliolids (0 to 3.5 %), small benthic foraminifera (< 0.5 %), echinoids (0 to 2.5 %), bryozoa (0 to 3.5 %) and undifferentiated bioclast debris (**Figure 3.17d**). Bioclasts may be concentrated in millimetre-scale vertical and sub-horizontal burrows. Most fossils are well preserved, although miliolids have an abraded appearance.

Fine-grained, green-brown matrix comprises up to 80 % of the coral wacke/floatstone facies (**Figures 3.11c** and **d**). The matrix is composed of micrite (48.5 to 62 %) and non-carbonate material (12 to 34 wt. %). Non-carbonate material is largely clay grade, although rare (< 2 %) silt-grade quartz grains are evident.

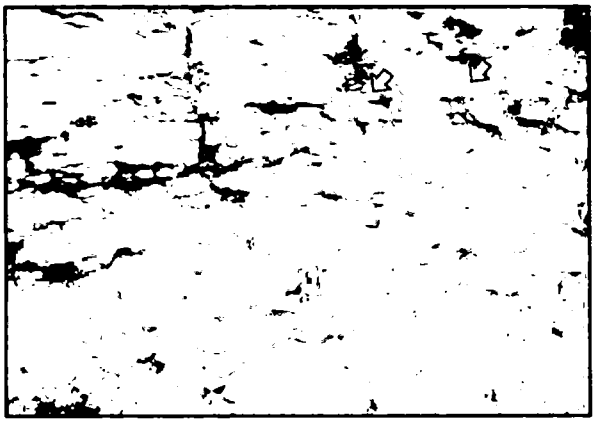


Figure 3.17a. Outcrop of the coral wacke/floatstone facies (log CA-7. bed 7). This facies demonstrates a nodular weathered appearance. Nodules are often aligned (arrowed), with the alignment a relict of bedding. Nodules are surrounded by blue-grey marl. It is postulated that the original depositional texture of this facies was coralline limestones interbedded with siliciclasitc rich marls. Scale bar=50cm

Figure 3.17b *In situ* delicate branching coral (log CA-7, bed 7 as shown in the above photograph). Pencil = 14 cm

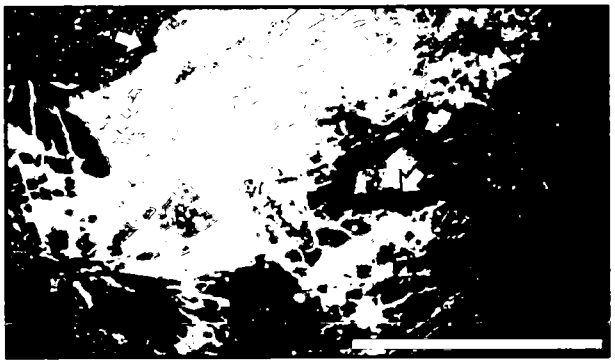


Figure 3.17c Photomicrograph (PPL) of the coral wacke/floatstone facies, illustrating a coral colony with a thin coralline algae crust (arrowed). Note the dark brown, homogenous matrix (sample LCA 37, log CA-4a, bed 1). Scale bar=5mm

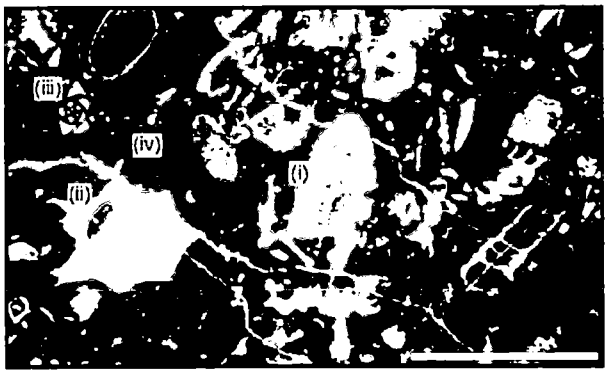


Figure 3.17d Photomicrograph (PPL) of the coral wacke/floatstone facies. This facies contains a modest diversity of biota, including fenestrate bryozoa (i), *Cerithium* gastropods (ii) and miliolids (iii) situated within a dark brown clay-rich micrite matrix (iv) (sample LCA 50, log CA-4 bed 11). Scale bar=2 mm

Diagenesis

Local recrystallisation of micrite into microspar is evident in most samples of this facies. All originally aragonitic biota have been leached with the resultant mouldic macropores completely cemented with inclusion-free, coarse drusy calcite cement (**Figures 3.17b** and **c**). Echinoid spines often show poorly developed syntaxial overgrowths. The nodular weathering is attributed to cementation of this facies, although it is has not possible to differentiate intergranular cements from the matrix.

Interpretation: depositional environment

Alignment of nodules is interpreted to be a remnant of depositional bedding. It is postulated from the preservation of bedding that the original depositional texture of this facies was coralline limestones interbedded with non-fossiliferous, siliciclastic rich marls. Deposition under normal marine conditions within the photic zone is interpreted from the presence of *in situ* coral colonies (Hallock and Glenn 1985, 1986). Coral colonies are sparsely distributed and small in comparison to the coral mixstone and coral domestone facies.

Low-energy conditions are inferred from the presence of delicate coral colonies. This is consistent with the fine-grained matrix. It is concluded from the high non-carbonate component of samples that this environment experienced a significant siliciclastic input. The gastropod *Cerithium* is described from a number of brackish settings in the Eocene of the Nummulitique (Pairis and Pairis 1975, Sayer 1995). It is postulated that inputs of freshwater may have accompanied siliciclastic input, limiting the growth potential of corals and other stenohaline biota. The thin, laterally extensive laminar growth form of coralline algae is consistent with low-energy settings where competition for space is limited (Minnery *et al.* 1985, Minnery 1990).

Clay to silt-grade siliciclastic grains would have been deposited from suspension. Branching corals are relatively common in settings where sedimentation rates are high, as steep upright faces are less susceptible to sediment settling (Grasso and Pedley 1988). Steep surfaces will experience a lower degree of incident radiation, thus would not be advantageous in turbid settings. It is suggested that turbidity was within tolerable levels as sediment would have settled quickly from suspension in a low energy setting, and clays may flocculate thus settle from suspension at higher water energies. Relatively high turbidity is suspected for beds where platy *Cyathoseris* corals developed rather than branching corals (i.e. log CA-7, bed 16).

It is postulated that the delicate, branching coral colonies grew constratally i.e. within the sediment with only a small proportion of the colony above the substratum. A constratal growth fabric is one in which the vertical organic growth occurred at a similar rate to sediment accumulation (Insalaco 1998). Constratal growth is supported by the relative absence of a cryptic community that was well developed in the coral mixstone and coral domestone facies (**Sections 3.3.1.1** and **3.3.1.2**).

In summary, the coral wacke/floatstone facies represents patchy coral colonisation within a low-energy marine setting with a significant siliciclastic input. Conditions within this environment favoured a limited biota in comparison to other carbonate-dominated facies in the Calders study area. Similar facies are encountered within coeval carbonate sediments on the northern margin of the Vic Basin (Alvarez 1991, Franques-Faixa 1996). Platy *Cyathoseris* corals developed directly on the slope of delta lobe deposits, where incident light was low and horizontal growth was favoured over vertical growth due to shallow water depths (Alvarez 1995).

3.3.1.5 Coralgal foraminifera rudstone

Lithologies: Coralgal foraminifera rudstone
Coralgal foraminfera rhodolithic rudstone

Occurrence and bed characteristics

This facies is encountered within carbonate intervals 1 to 6 (**Figure 3.9**) and often occurs associated with the coralgal foraminifera float/rudstone and foralgal grainstone facies. The coralgal foraminifera rudstone facies have a nodular weathered appearance. Nodules are up to 15 cm in diameter and often include a colonial coral colony. Beds dip consistently to the northwest. Nodules are surrounded by blue-grey non-carbonate mud. Measured bed thickness ranges from 1.2 to 2.4 m. Upper and lower bed contacts are sharp but non-erosional. This facies can be traced laterally up to 500 m.

Lithological description

Exposed surfaces of this facies have a crystalline texture with a buff to light grey colouration. Fresh surfaces are a light blue-grey colour with localised pink patches. The coralgal foraminifera rudstone facies is characterised by *in situ,* toppled and fragmented colonial corals with a rich encrusting community, situated within a bioclastic rudstone matrix. The total coral skeletal volume of this facies ranges from

20 to 40 %, and colonies do not form a framework. Coral colonies are up to 50 cm and demonstrate branching, dome and foliaceous growth morphologies. Coral genera identified include *Acropora, Porites, Actinacis* and *Favites*, although more genera are likely to be present. Coralline algae, foraminifera and serpulids encrust *in situ* coral colonies (**Figure 3.18a**).

Foralgal rhodoliths, comprising up to 8.5 % of this facies, have tight, concentric laminar growth forms, often with thick columnar branches up to 0.5 cm in length. Rhodoliths are up to 6 cm in diameter and nucleated on coral fragments (**Figure 3.19a**). Foraminifera present within rhodoliths are *Gypsina*, *Fabiania* and *Haddonia*. Laminar *Gypsina* nodules up to 1 cm in diameter are also present in this facies (**Figure 3.18b**).

The rudstone matrix contains abundant angular coral fragments (4 to 25 %) up to 10 cm in length (**Figure 3.19b**). Laminar coralline algae fragments (1.5 to 25 %), detached from rhodoliths and corals, are up to 4 mm in length. The coralgal foraminifera rudstone facies contains a diverse benthic foraminifera assemblage that includes *Amphistegina* (0 to 7.5 %), *Gypsina* (0 to 4 %), *Nummulites* (0 to 2%), *Calcarina* (0 to 3 %), miliolids (0 to 5%) and discorbids (0 to 4.5 %). Large encrusting forms (*Fabiania, Haddonia*, *Chapmanina* and victoriellids) comprise up 15 % of this facies. Foraminfera are preserved intact and largely unabraded. *Amphistegina* and *Nummulites* tests demonstrate robust, lenticular morphologies with thick test walls. Test diameter ranges from 1 to 5 mm.

Other biota present are bryozoa (0 to 4 %), echinoids (0 to 4.5 %), gastropods (0.5 to 3.5 %) and undifferentiated fine-grained bioclastic debris (**Figures 3.19b** and **c**). Most fossils are fragmented and poorly sorted.

Thirty two to 65 % of this facies is composed of dark brown micrite. The total non-carbonate component ranges from 5 to 21.5 wt. %. Non-carbonate grains are clay to very fine silt grade, and cannot be differentiated from the micrite (**Figures 3.19a** and **b**).

Figure 3.18a Photomicrograph (PPL) of the coralgal foraminfera rudstone facies. Coral fragments (i) are encrusted by laminar coralline algae (ii) and acervulinid foraminfera (iii) (sample LCA 53, log CA-4 bed 7). Scale bar=2 mm.

Figure 3.18b Photomicrograph (PPL) of the coralgal foraminfera rudstone facies. Laminar *Gypsina* nodule (i) that nucleated around matrix or sea grass, which later decayed. Texturaliid foraminfiera (ii) are also present in this facies (sample LCA 54, log CA-4 bed 5). Scale bar=2 mm.

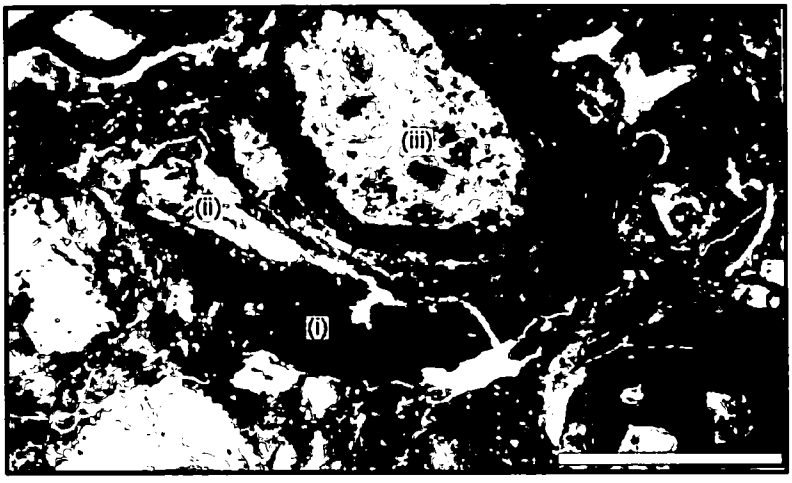


Figure 3.19a Photomicrograph (PPL) of the coralgal foraminfera rudstone facies. Thick coralline algae (i) and *Haddonia* (ii) crusts developed around coral fragments (iii) (Sample LCA 53, log CA-4b). Scale bar=2mm.

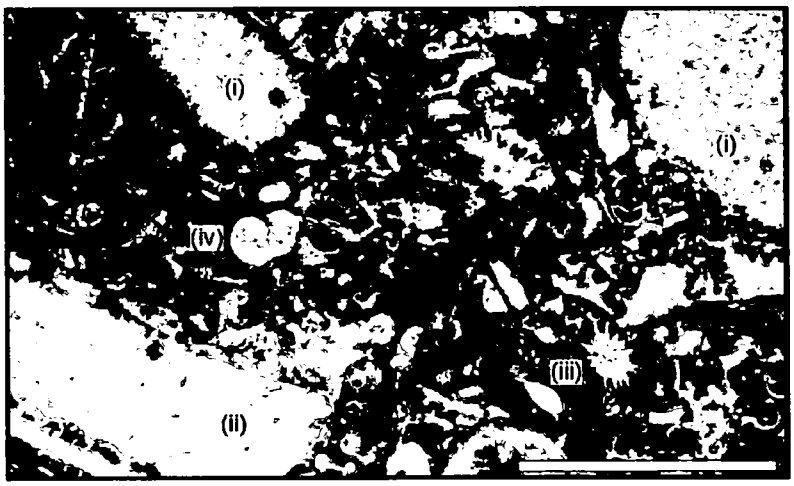


Figure 3.19b Photomicrograph (PPL) of the coralgal foraminfera rudstone facies. Replacement calcite spar may be inclusion rich (i) or relatively clear (ii). Other foraminfera present include *Calcarina* (iii) and discorbids (iv) (sample LCA 99e, log CA-11 bed 1). Scale bar=2mm.

Figure 3.19c Photomicrograph (PPL) of the coralgal foraminfera rudstone facies. Coral fragment (i) demonstrating multiple thick coralline algae and *Gypsina* crusts. Fragmented and abraded echinoid plates (ii) are present in the matrix. The matrix of this facies is composed predominantly of micrite that shows local recrystallisation to microspar (sample LCA 51, log CA-4b).

Diagenesis

The micrite matrix has undergone patchy recrystallisation (**Figure 3.19c**). Intragranular microspar (recrystallised matrix) is observed within foraminifera tests, and echinoid spines demonstrate syntaxial calcite overgrowth cements. The original skeletons of corals and molluscan fragments have been leached with the resultant biomoulds occluded with equant and drusy calcite spar cement (**Figure 3.19b** and **c**). Centimetre-scale dissolution vugs partially in-filled with coarse blocky calcite are observed in the field. Blue-grey clay partings observed in outcrop are interpreted as macro-dissolution seams that developed late in the diagenetic history as a consequence of compaction.

Interpretation: depositional environment

The coralgal foraminifera rudstone facies contains a marine faunal assemblage. Normal marine conditions are interpreted from the diversity of stenohaline biota and the abundance and diversity of *in situ* coral colonies (Ghose 1977, Hallock and Glenn 1985, Hohenegger *et al.* 1999, Geel 2000).

Deposition within the photic zone is inferred from the presence of symbiont-bearing larger foraminifera such as *Amphistegina*, *Calcarina* and *Nummulites* (Leutenegger 1984, Hallock and Glenn 1985, 1986). The robust lenticular morphology of *Amphistegina* and *Nummulites* is indicative of the shallow to intermediate part of their depth range in moderate to high-energy, agitated conditions (Hottinger 1983, Hallock and Glenn 1985). Agitation is confirmed by the presence of tight, laminar foralgal rhodoliths (Bosence 1983). *Gypsina, Amphistegina* and *Calcarina* adapt to an epiphytic lifestyle, encrusting sea grass and macroalgae (Hohenegger *et al.* 1999).

There is no evidence to suggest the coralgal foraminifera rudstone facies possessed any topographic relief. It is inferred that the growth type tended towards suprastratal (in the sense of Insalaco 1998), with coral colonies extending vertically beyond the rudstone substrate. This is supported by the diverse, flourishing cryptic community (e.g. *Haddonia, Chapmanina,* victoriellids etc). It is postulated that the vertical development (and topographic relief) may have limited by the lack of accommodation space i.e. the water was too shallow.

A moderate siliciclastic input is indicated from the significant percentage of non-carbonate material in the matrix of this facies. Low-energy periods are inferred

from the fine grainsize of siliciclastics that would have been deposited from suspension. Alternatively, *in situ* corals that stand proud of the substrate may influence and modify the hydrodynamic conditions trapping sediment in their immediate vicinity (Insalaco 1998). Suprastratal growth fabrics have a high potential for trapping sediments from suspension (Insalaco 1998).

In summary, the coralgal foraminifera rudstone facies is interpreted as coral debris with patchy *in situ* coral development within a moderate to high-energy fore- or back-reef setting. This facies is comparable to the Coralgal Packstone Bindstone and Algal Bindstone facies of Sayer (1995), which are interpreted as patch reefs and algal shoals that developed within an agitated inner-ramp setting. Repeated overgrowth of corals by coralline algae as a consequence of changes in incident light (through variations in water depth and turbidity) is described from the Early Eocene Sabassona Reef situated on the northern margin of the Vic Basin (Taberner and Bosence 1985).

3.3.1.6 Coralgal foraminifera float/rudstone

Occurrence and bed characteristics

The coralgal foraminifera float/rudstone facies is encountered within carbonate intervals 1 to 6 (**Figure 3.9**). This facies is often interbedded with the coralgal foraminifera rudstone and coral mixstone facies.

This facies demonstrates a nodular weathered appearance with blue-grey clay partings that extend laterally less than 1 m. Nodules up to 10 cm in length may be aligned and often include a coral colony. Measured bed thickness ranges from 0.5 to 1.5 metres. This facies can be traced laterally for several hundred metres.

Lithological description

Exposed surfaces of the coralgal foraminifera float/rudstone facies have a crystalline texture and weather to buff to light grey colour. Fresh surfaces are a light blue-grey colour with localised pink patches. This facies is characterised by sparsely distributed *in situ* and overturned colonial corals situated within a bioclastic float/rudstone matrix. The total *in situ* coral skeletal volume is less than 20 %. *In situ* and toppled corals are mainly branching forms with rare domal and foliaceous colonies. Coral genera identified are *Porites, Actinacis, Cereiphyllia, Acropora* and *Stylophora.*

The float/rudstone matrix contains a diverse biota including coral fragments (0.5 to 48.5 %) that are encrusted by coralline algae and foraminifera such as *Fabiania* (**Figures 3.20a** and **b**). Coral fragments are up to 5 cm and may act as nuclei for rhodoliths (**Figure 3.20c**) that comprise up to 10 % of this facies. Coralline algae crusts are less than 1 mm thick. Fragments of laminar coralline algae up to 0.5 cm are quite common in the matrix, comprising 3 to 21 % of this facies. In addition, coralline algae peloids up to 0.5 mm comprise up to 2 % of bioclasts within this facies (**Figures 3.21a** and **b**).

This facies contains a diverse foraminifera assemblage (**Figures 3.21a** and **b**) including miliolids (0 to 2.5 %), *Gypsina* (0 to 5 %), *Amphistegina* (< 0.5 %) and *Nummulites* (0 to 2.5 %). Large encrusting foraminifera such as *Haddonia, Fabiania* and victoriellids comprise 1.5 to 9 % of this facies. *Gypsina* occurs as laminar forms up to 2 mm in length. Miliolid tests are highly abraded and may be slightly reddened (**Figure 3.21b**). *Amphistegina* and *Nummulites* tests are less than 2 mm in diameter, and demonstrate robust, lenticular morphologies with thick test walls.

Minor components of the coralgal foraminifera float/rudstone facies are echinoids (0 to 2.5 %), *Cerithium* gastropods (0 to 6.5 %), bivalves (< 0.5 %), sponge spicules (<0.5 %) and fenestrate bryozoa (0 to 5.5 %). Most fossils are fragmented, although *Cerithium* gastropods up to 20 cm in length are preserved intact (**Figure 21c**). In summary, most fossils in the coralgal foraminifera float/rudstone facies are fragmented and poorly sorted.

Micrite comprises up to 50 % of this facies. Micrite is dark brown to grey colour in thin section. The total non-carbonate content of this facies ranges from 8 to 12.5 wt. %. Non-carbonate grains are clay-grade and cannot be differentiated from micrite.

Diagenesis

Patchy recrystallisation of the micrite is evident, with samples of this facies containing 1 to 9.5 % replacement microspar. Intragranular calcite microspar cement is observed within foraminifera tests, and echinoid spines have well-developed syntaxial calcite overgrowth cements. Originally aragonitic organisms have been leached with the mouldic pores later occluded with inclusion-free drusy and equant calcite cement (**Figures 3.20a** and **b**).

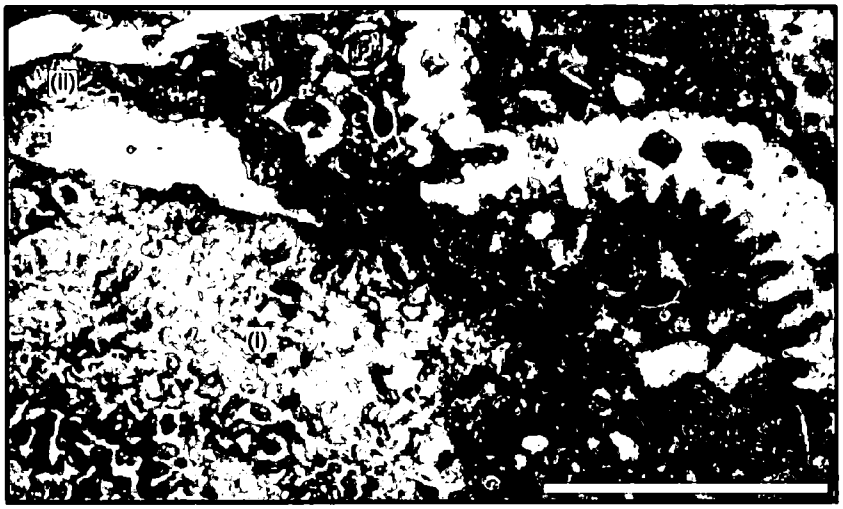


Figure 3.20a Photomicrograph (PPL) of the coralgal foraminifera float/rudstone facies. Corals such as *Actinacis* (i) are encrusted by larger foraminfera such as *Fabiania* (ii) (sample LCA 18, log CA-3 bed 3). Scale bar=2mm.

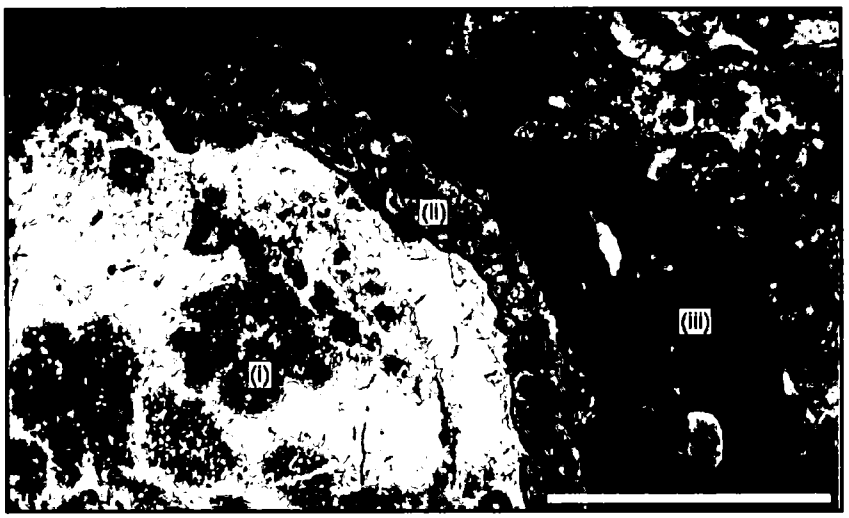


Figure 3.20b Photomicrograph (PPL) of the coralgal foraminifera float/rudstone facies. Fragment of *Stylophora* (i) encrusted by *Fabiania* (ii) and coralline algae (iii) (sample LCA 27, log CA-3 bed 4). Scale bar=2 mm.

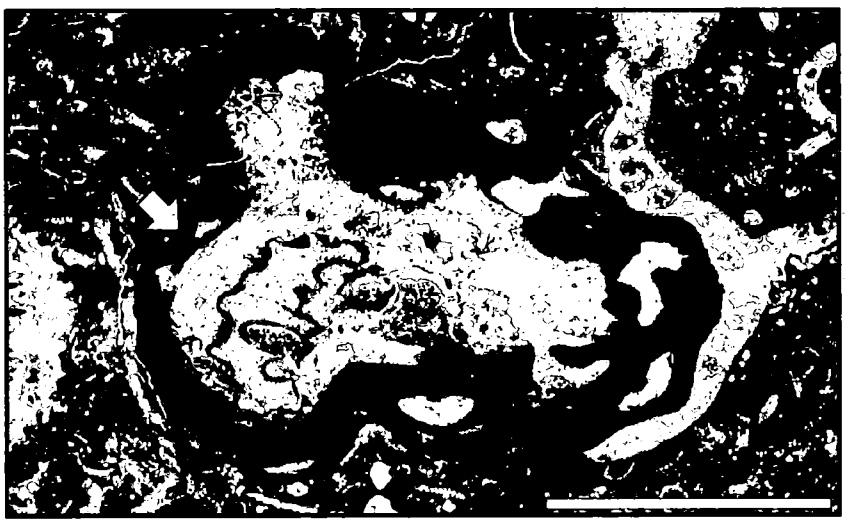


Figure 3.20c (PPL) Irregular, nodular rhodolith (arrowed) within the coralgal foraminifera float/rudstone (sample LCA 27, log CA-3 bed 4). Scale bar=2 mm.

Figure 3.21a Photomicrograph (PPL) of the coralgal foraminfera float/rudstone facies. The matrix contains a diverse faunal assemblage including laminar *Gypsina* (i), *Haddonia* (ii), abraded coralline algae (iii) and coral fragments (iv) (sample LCA 45, log CA-4). Scale bar=2 mm.

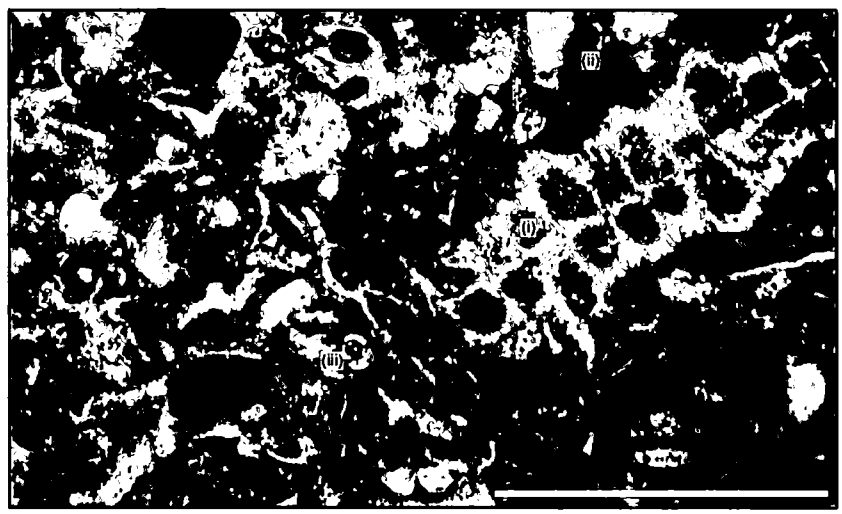


Figure 3.21b Photomicrograph (PPL) of the coralgal foraminfera rudstone facies. Fossils in the matrix are typically fragmented and include fenestrate bryozoa (i), peloidal coralline algae (ii) and miliolids (iii) (sample LCA 43a, log CA-4). Scale bar=2 mm.

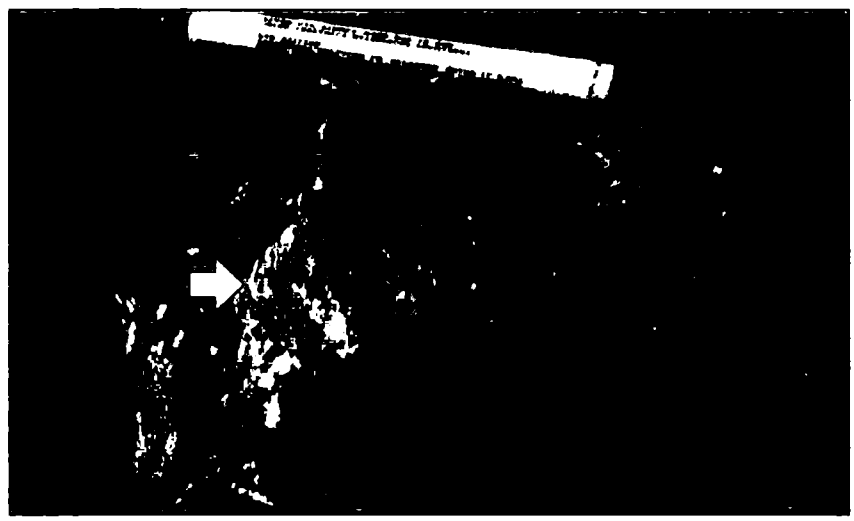


Figure 3.21c Large *Cerithium* gastropod (arrowed) within the coralgal foraminifera float/rudstone facies (log CA-4a, bed 14). Pen=15cm

Interpretation: depositional environment

The coralgal foraminifera float/rudstone facies contains a marine faunal assemblage. Normal marine conditions are inferred from the diversity of stenohaline biota. Deposition within the photic zone is inferred from the occurrence of *in situ* coral colonies and well-preserved symbiont-bearing larger benthic foraminifera (i.e. *Amphistegina* and *Nummulites*) (Ghose 1977, Hallock and Glenn 1985, Hohenegger *et al.* 1999, Geel 2000).

There is no evidence to suggest the coralgal foraminifera float/rudstone facies possessed any topographic relief. Suprastratal coral growth, with rare colonies extending vertically beyond the float/rudstone substrate, is inferred from the diverse, flourishing cryptic community (e.g. *Haddonia, Fabiania*). Suprastratal growth fabrics have a high potential for trapping sediments from suspension (Insalaco 1998). Corals that stand proud of the substrate may influence and modify the hydrodynamic conditions trapping suspended sediment in their immediate vicinity (Insalaco 1998). Due to the patchy nature of coral colonisation in this facies, it is postulated that an overall low-energy hydrodynamic regime aided the deposition of fine-grained sediments from suspension.

In addition to the factors described above, slight variations in the hydrodynamic regime have also influenced the texture of this facies. The small, robust morphology of *Amphistegina* is characteristic of the shallower parts of its depth range in agitated waters (Leutenegger 1984, Hallock and Glenn 1985). An abundance of *Gypsina* and miliolids may also be characteristic of shallow, high-energy, agitated environments (Chaproniere 1975, Reid and MacIntyre 1988, Hallock 1998, Murray 1991, Geel 2000). It is suggested that these relatively high-energy fauna have been transported into this environment during periods of high-energy. This is consistent with the fragmented nature of most fossils in this facies, although the limited abrasion demonstrates bioclasts have not been transported large distances.

In summary, the coralgal foraminifera float/rudstone facies is interpreted as deposits of patchy coral development within a low-energy environment experiencing a moderate siliciclastic sediment input. From its common association, it is postulated that this facies represents the distal or partially protected equivalent of the coralgal foraminifera rudstone facies (**Section 3.3.1.5**).

3.3.1.7 Foralgal pack/grainstone

Lithologies: Foralgal packstone
Foralgal pack/grainstone
Foralgal pack/grainstone
Foralgal rhodolith grainstone

Occurrence and bed characteristics

The foralgal pack/grainstone facies is a very common facies and is encountered in carbonate intervals 2 to 6 (**Figure 3.9**). This facies is characterised by massive beds that locally demonstrate nodular weathering. Measured bed thickness is up to 2 m. Beds can be traced laterally (along strike) for over 1 km. Upper and lower bedding contacts are sharp but non-erosional. This facies is often intercalated with the *Gypsina* grainstone, *Gypsina* arenite and the miliolid coralline algae pack/grainstone facies.

Lithological description

The exposed surface of the foralgal grainstone facies has a crystalline texture and weathers to a pale brown-grey colour. Fresh surfaces are a blue-grey colour with localised pink-grey patches.

The foralgal pack/grainstone facies is characterised by sparsely distributed *in situ* branching and domal coral colonies situated within a bioclastic pack/grainstone. The total *in situ* coral skeletal volume ranges from 20 to 25 %. Coral colony size ranges from 3 to 15 cm. Coral genera identified are *Porites, Acropora, Cereiphyllia* and *Favites*. Coralline alage, foraminifera and bryozoa encrust corals (**Figure 3.22**). Coral fragments up to 7 cm in length comprise 1 to 15 % of this facies.

Foraminifera and fragmented coralline algae dominate the foralgal pack/grainstone facies. Coralline algae occur as fragmented and abraded laminar forms up to 2 mm in length (6.5 to 21.5 %), sand-grade peloids (up to 20 %) and articulated forms (*Corallina*) (**Figure 3.23a**). Foralgal rhodoliths up to 7.5 cm in diameter are common in this facies (**Figure 3.23b**). Rhodoliths are subspherical to elliptical with tight, concentric laminar crusts. Branching rhodoliths are rare. Where present, branches are columnar and <5 mm long.

The foralgal pack/grainstone facies contains a diverse foraminifera assemblage that includes miliolids (0.5 to 3 %), texturaliids (0.5 to 3.5 %), *Gypsina* (1 to 10 %),

Calcarina (0 to 2 %), *Rotorbinella* (0 to 9.5 %), *Amphistegina* (0 to 2.5 %), *Nummulites* (0 to 1 %) and *Discocyclina* (**Figures 3.23c** and **3.24b** and **c**). Large encrusting foraminifera (*Fabiania, Haddonia, Chapmanina* and victoriellids) comprise up to 9.5 % of this facies. Miliolids (up to 0.75 mm in diameter) are preserved intact but have an abraded appearance. *Gypsina* tests, up to 3.0 mm in length, may show poorly developed micrite envelopes (**Figure 3.23c**). *Amphistegina* and *Nummulites* are small (typically less than 5 mm) with robust lenticular test morphologies. *Discocyclina* tests are up to 1 cm in diameter, and demonstrate flattened discoidal test morphologies (**Figure 3.24b**). Spinose *Calcarina* tests, up to 2 mm in diameter, are preserved intact within samples containing a higher percentage of micrite (**Figure 3.24a**).

Subordinate bioclasts represented in this facies are bryozoa (0.5 to 1 %), echinoids (0 to 6.5 %), molluscs (0.5 to 1 %) and serpulids (0 to 1 %). Dark brown, mixed carbonate-siliciclastic mud comprises 1.5 to 18 % of this facies. The total non-carbonate content of this facies ranges from 5.5 to 14.5 wt. %. Siliciclastic grains are clay to medium sand grade (**Figure 3.22a**).

Diagenesis

Intergranular areas have been cemented with equant calcite spar (**Figure 3.23b**). Intergranular calcite spar cement comprises up to 59 % of this facies. Micrite has locally recrystallised into microspar. Originally aragonitic fossils have been leached with the resultant mouldic macropores completely occluded with clear, drusy calcite cement. Pores within foraminifera chambers have also been occluded with calcite (**Figure 3.23b**). Syntaxial calcite overgrowth cements have developed on echinoid spines.

Interpretation: depositional environment

The foralgal pack/grainstone facies contains a marine faunal assemblage. Normal, open marine conditions are inferred from the diversity of stenohaline biota.

Deposition within the photic zone is inferred from the presence of *in situ* corals and symbiont-bearing larger foraminifera such as *Calcarina, Amphistegina* and *Nummulites* (Leutenegger 1984, Hallock and Glenn 1985, 1986, Hohenegger *et al.* 1999). Is inferred from the diverse cryptic community preserved in this facies that corals were growing suprastratally (in the sense of Insalaco 1998). Large encrusting

foraminifera (such as *Haddonia, Chapmanina* and *Fabiania*) inhabited cryptic environments within coral debris (Franqués-Faixa 1996, Romero 2001, Romero *et al.* 2002). The presence of sea floor vegetation (sea grass) is inferred from the abundance of epiphytic foraminifera (*Gypsina* and *Calcarina*) and articulated coralline algae (Chaproniere 1975, Ghose 1977, Carbone *et al.* 1994, Travé 1996, Hohenegger *et al.* 1999).

The robust lenticular morphology of *Amphistegina* and *Nummulites* is indicative of the shallow to intermediate part of their depth range in moderate to high-energy conditions (Hottinger 1983, Hohenegger 1984, Hallock and Glenn 1986). An abundance of *Amphistegina* and *Gypsina* is an indicator of current-swept conditions (Hallock and Glenn 1985, 1986, Li and Jones 1997, Li *et al.* 1997, Hohenegger *et al.* 1999). This is consistent with the abundance of abraded miliolids and the highly abraded nature of most bioclasts in this facies (Chaproniere 1975, Ghose 1977, Geel 2000). It is postulated that samples containing a more significant proportion of micrite (i.e. LCA 231, log CA-15) were deposited within a relatively lower-energy setting. This is consistent with the well-preserved spinose *Calcarina* tests (**Figure 3.24a**).

In summary, it is concluded that the foralgal pack/grainstone facies formed within an agitated setting with open marine circulation in the shallow part of the photic zone. This facies has affinities with the Algae Debris Packstone facies of Sayer (1995), interpreted as a fore-bank wash-over shoal deposit. The foralgal pack/grainstone facies is also comparable with bioclastic pack/rudstone sediments described from the Eocene Qum Formation (Iran) that are interpreted as shallow water shoals, although more restricted conditions are indicated from the absence of corals (Okhravi and Amini 1998).

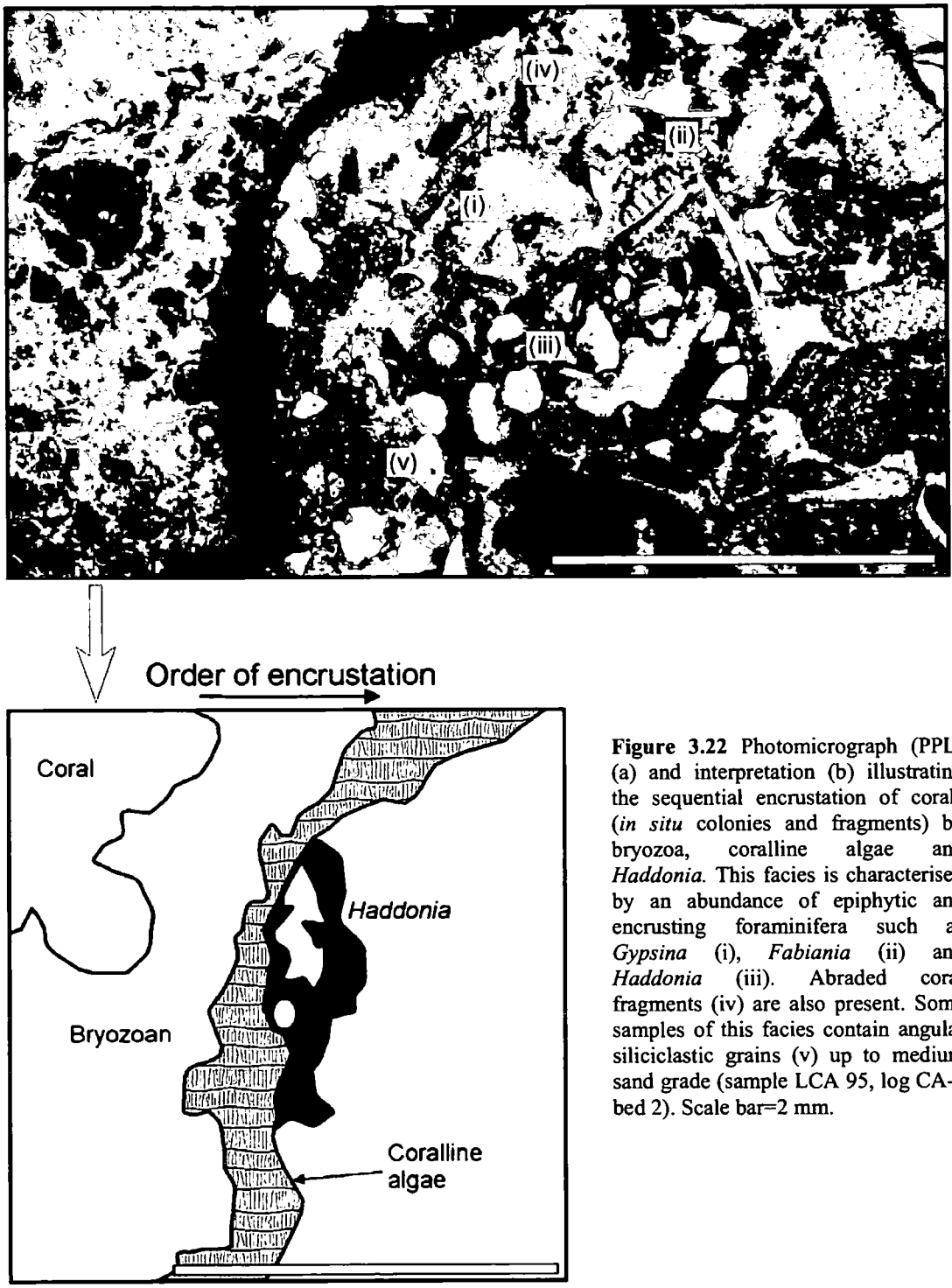


Figure 3.22 Photomicrograph (PPL) (a) and interpretation (b) illustrating the sequential encrustation of corals (*in situ* colonies and fragments) by bryozoa, coralline algae and *Haddonia*. This facies is characterised by an abundance of epiphytic and encrusting foraminifera such as *Gypsina* (i), *Fabiania* (ii) and *Haddonia* (iii). Abraded coral fragments (iv) are also present. Some samples of this facies contain angular siliciclastic grains (v) up to medium sand grade (sample LCA 95, log CA-9 bed 2). Scale bar=2 mm.

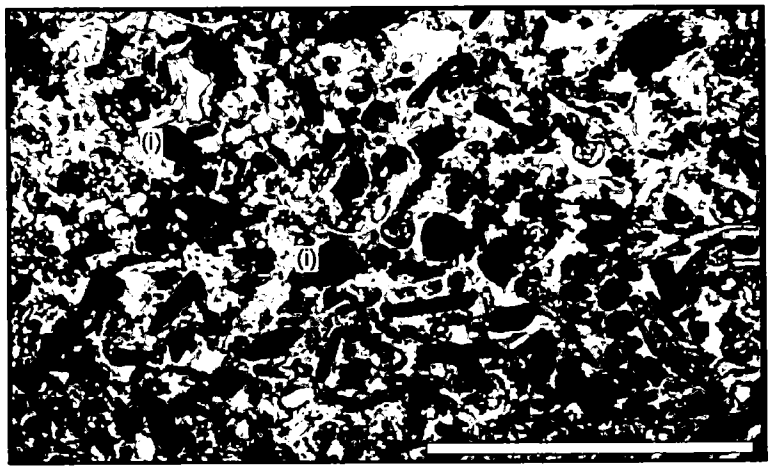


Figure 3.23a Photomicrograph (PPL) of the foralgal pack/grainstone facies. This facies is characterized by an abundance of abraded coralline algae fragments (i) situated within a partially winnowed matrix (sample LCA 37, log CA-3 bed 17). Scale bar=5 mm.

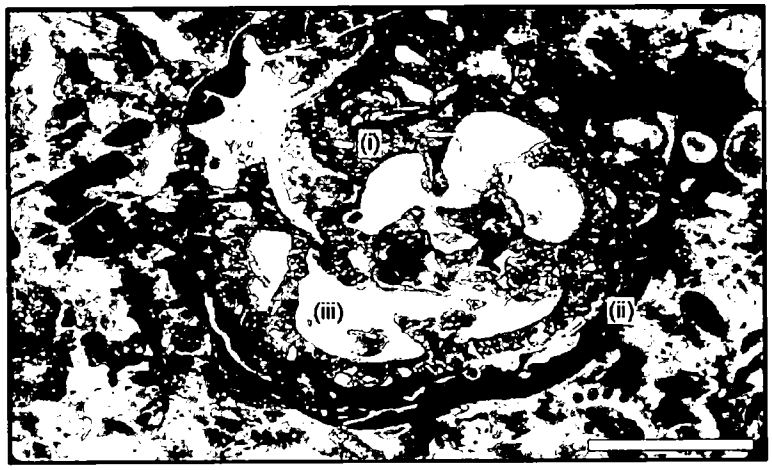


Figure 3.23b (PPL) Foralgal rhodolith composed of *Haddonia* (i) and laminar coralline algae (ii). Note the clear drusy calcite cement (iii) within *Haddonia* chambers (sample LCA 41, log CA-4a). Scale bar=5 mm.

Figure 3.23c (PPL) Abundant *Gypsina* tests (i) associated with serpulids (ii). An epiphytal mode of life is inferred from the rounded, elongated morphology of *Gypsina*. Note the dark outlines of foraminifera tests, interpreted as incipient micrite envelope development (sample LCA 87a, log CA-4b). Scale bar=2 mm.

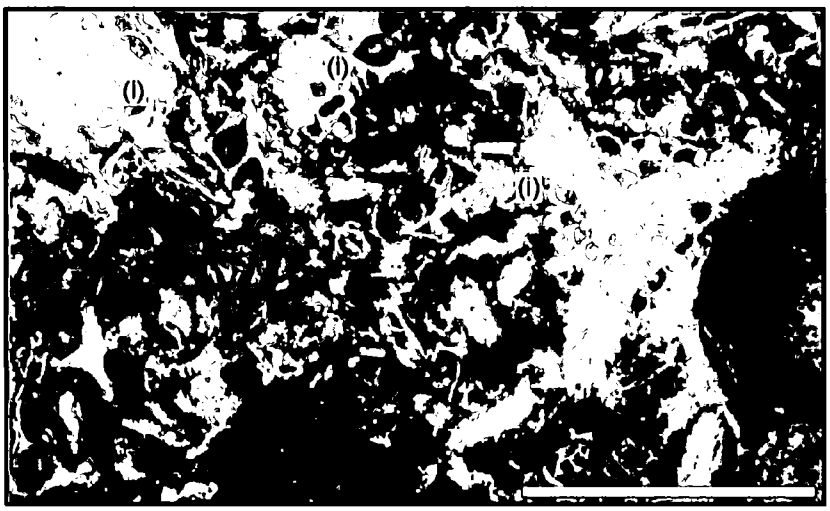


Figure 3.24a Photomicrograph (PPL) of the foralgal pack/grainstone facies. The symbiont-bearing foram *Calcarina* is common in some samples of this facies (sample LCA 231, log CA-15 bed 9). Scale bar=2 mm.

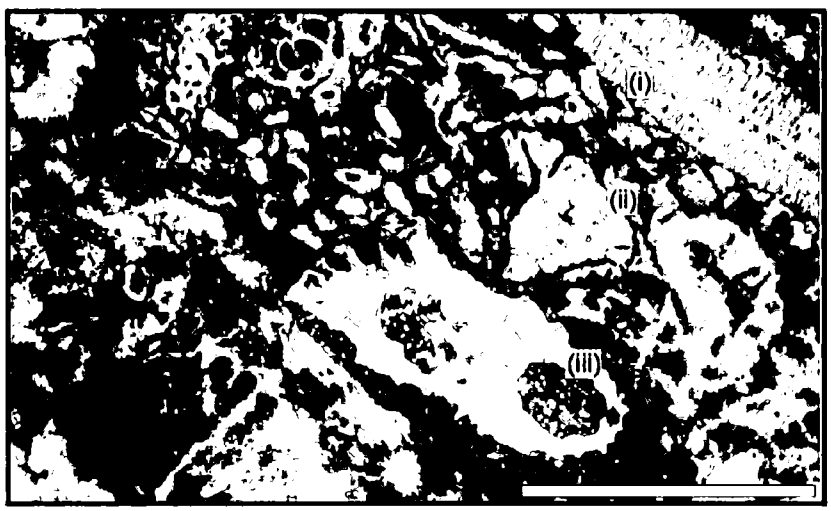


Figure 3.24b Photomicrograph (PPL) of the foralgal pack/grainstone facies. Larger benthic forams present include *Discocyclina* (i) and victoriellids (ii). *Gypsina* is observed encrusting coral fragments (iii) (sample LCA 231, log CA-15 bed 9). Scale bar=2 mm.

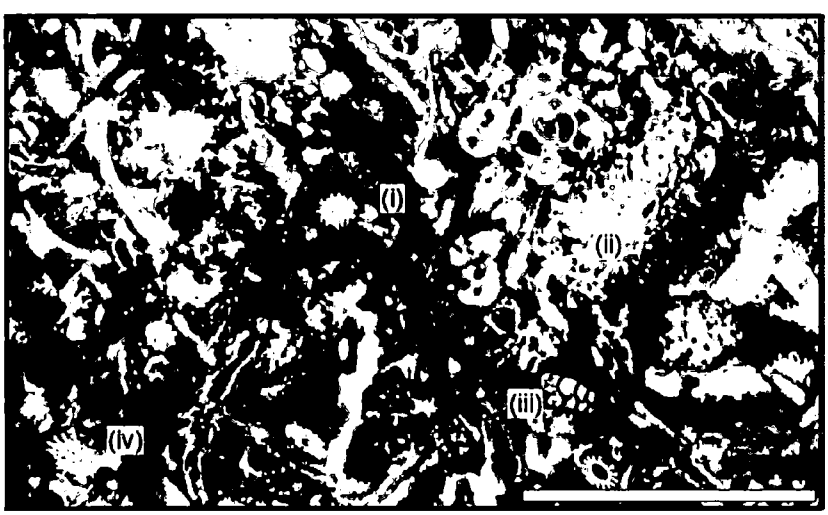


Figure 3.24c Photomicrograph (PPL) of the foralgal pack/grainstone facies. In addition to encrusting forms, articulated coralline algae (i) is common in this facies. Coral fragments are encrusted by coralline algae (ii). Texturaliid foraminifera (iii). Note the preservation of dark brown micrite (iv) (sample LCA 231, log CA-15 bed 9). Scale bar=2 mm.

3.3.1.8 *Gypsina* grainstone

Lithologies: *Gypsina* grainstone
Gypsina pack/grainstone

Occurrence and bed characteristics

This facies is encountered throughout the studied section but is particularly abundant in the upper half of the succession (carbonate intervals 4 to 6). Measured bed thickness ranges from 1.4 to 1.8 m. Bedding contacts are sharp but non-erosional, and beds can be traced laterally along strike for at least 1 km. Units are largely massive, with a locally developed nodular weathered appearance.

Lithological description

Exposed surfaces of the *Gypsina* grainstone facies weather to a dark grey colour. Fresh surfaces are a pale grey colour. Up to 25 % of this facies is composed of intact and fragmented laminar *Gypsina* tests (**Figure 3.25a**). Tests are up to 5 mm in length and may have a poorly developed micrite envelope. Other foraminifera present in this facies are miliolids (2 to 2.5 %), texturaliids (0 to 5 %), *Calcarina* (0 to 6 %), *Amphistegina* (0 to 8 %), *Rotorbinella* (0 to 3.5 %) and *Nummulites* (< 0.5 %). Abraded miliolids and texturaliids are up to 1.5 mm in diameter. *Calcarina* is typically fragmented, although rare intact specimens up to 2 mm in diameter are present. *Amphistegina* and *Nummulites* tests have a robust lenticular morphology with thick walls. Tests are typically fragmented.

Fabiania and *Haddonia* are typical constituents of spheroidal, concentric laminar foralgal rhodoliths, although detached tests are also observed (**Figure 3.25b**). Rhodoliths are up to 2 cm in diameter and are often fragmented. Rhodolith nuclei are coral fragments. Coralline algae also occur as detached and fragmented laminar crusts (1 to 9 %) and articulated forms (< 1 %). Fragments show moderate abrasion. Other fossils present in this facies are corals (0 to 12.5 %), echinoids (1 to 5 %), molluscs (< 1 %), bryozoa (0 to 1.5 %) and serpulids (< 1 %) (**Figure 3.25b**). Fossils are fragmented and poorly sorted, and show evidence of abrasion.

The matrix of the *Gypsina* grainstone facies is texturally variable, comprising microspar with localised dark brown mixed carbonate-siliciclastic mud patches and skeletal hash. The non-carbonate proportion of this facies ranges from 7.5 to

Figure 3.25a Photomicrograph (PPL) of the *Gypsina* grainstone facies. This facies is dominated by *Gypsina* (i) with subordinate *Rotorbinella* (ii). The contacts between foraminifera may be dissolutional (arrowed). Clays occur concentrated along dissolution seams (sample LCA 181, log CA-12 bed 19). Scale bar=2 mm.

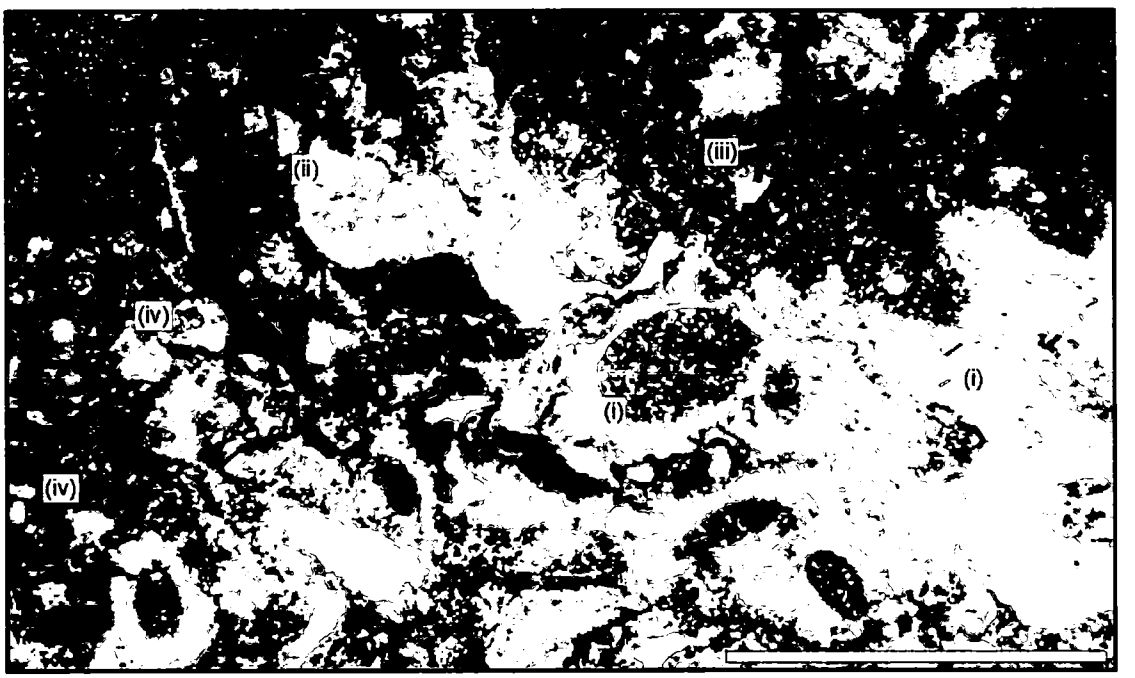


Figure 3.25b Photomicrograph (PPL) of the *Gypsina* grainstone facies. Subordinate biota present in this facies include corals (i), *Rotorbinella* (ii), *Fabiania* (iii) and echinoids (iv). Corals are composed of a secondary drusy calcite cement. Also note the calcite cements within chambers of *Rotorbinella*. (sample LCA 98, log CA-9 bed 11). Scale bar=2 mm.

21.5 wt. %. Non-carbonate grains are clay to fine silt grade, and may be concentrated along dissolution seams (**Figure 3.25a**).

Diagenesis

Intragranular equant calcite cement is observed within foraminifera chambers (**Figure 3.25b**) and echinoid spines have well-developed syntaxial calcite overgrowth cements. Intergranular equant to locally drusy calcite comprises 23.5 to 30 % of this facies. Patchy recrystallisation of micrite into microspar is evident in some samples of this facies. Aragonitic fragments have been leached with the resultant porosity completely occluded with coarse, drusy calcite cement (**Figure 3.25b**). Post-depositional compaction of this sediment is evident from the dissolution seams, along which clays are concentrated (**Figure 3.25a**).

Interpretation: depositional environment

The *Gypsina* grainstone facies contains a marine faunal assemblage. Normal open marine conditions are interpreted from the abundance of stenohaline biota. Deposition within the photic zone is inferred from the presence of symbiont-bearing larger foraminifera such as *Calcarina, Nummulites* and *Amphistegina* (Leutenegger 1984, Hallock and Glenn 1985, 1986, Murray 1991, Hohenegger *et al.* 1999). The robust lenticular morphology of *Amphistegina* and *Nummulites* is indicative of the shallow to intermediate part of their habitat range in higher energy conditions (Hottinger 1983, Hallock and Glenn 1985, 1986). This is consistent with the presence of tightly concentric foralgal rhodoliths (Minnery *et al.* 1985, Reid and MacIntyre 1988, Minnery 1990) and the fragmented abraded bioclasts in the matrix. Large encrusting foraminifera such as *Fabiania* and *Haddonia* are derived from the rhodoliths, but also inhabited cryptic environments within coarse fossil debris (Romero 2001, Romero *et al.* 2002). The presence of sea grass is inferred from the abundance of *Gypsina* with miliolids, *Orbitolites* and *Calcarina* (Chaproniere 1975, Kitazato 1988, Murray 1991, Hohenegger *et al.* 1999). Articulated coralline algae are also known to have to have an epiphytic mode of life (Carbone *et al.* 1994).

Although grainstone textures imply high-energy conditions, the significant clay component suggests low-energy periods where sediment was deposited from suspension. It is postulated that the baffling effect of sea grass and flocculation of clays, encouraged the deposition of suspended sediment.

In summary, it is concluded that the *Gypsina* grainstone facies formed within an open marine, agitated shallow-water setting in the upper part of the photic zone. Fine-grained siliciclastic material was deposited from suspension during relatively quiet periods, and was trapped by sea floor vegetation.

3.3.1.9 Miliolid coralline algae pack/grainstone

Lithologies: Miliolid coralline algae pack/grainstone
Miliolid coralline algae packstone

Occurrence and bed characteristics

The miliolid coralline algae pack/grainstone facies is identified in carbonate intervals 5 and 6 only (**Figure 3.9**). This facies demonstrates a nodular weathered appearance. Nodules are up to 10 cm in length and may be aligned. Measured bed thickness is consistently around 1 m, and beds may be traced laterally up to 500 m. Lower bedding contacts are sharp but non-erosional. Upper contacts are gradational.

Lithological description

Exposed surfaces of the miliolid coralline algae pack/grainstone facies has a crystalline texture and weathers to a pale grey to buff colour. Fresh surfaces are a pale blue/grey colour with pink patches. The miliolid coralline algae facies is characterised by an abundance of miliolids (9.5 to 10.5 %), peloids (10 to 15 %) and coralline algae (up to 12.5 %) (**Figure 3.26a** and **b**). The miliolid assemblage characteristic of this facies is highly diverse. Miliolid tests up to 0.5 mm in diameter are reddened and abraded (**Figure 3.26b**). Peloids, interpreted as abraded coralline algae fragments, are up to 0.25 mm in diameter. Laminar coralline algae fragments are up to 0.5 mm in length (**Figure 3.26b**).

Subordinate biota present in the miliolid coralline algae pack/grainstone facies are fragmented corals (5 %), *Gypsina* (0 to 4.5 %), *Orbitolites* (< 0.5 %), texturaliids (0 to 2 %), echinoids (0 to 1 %), bryozoa (0 to 1.5 %) and molluscs (0.5 to 2 %). Encrusting foraminifera (*Chapmanina* and victoriellids) comprise up to 1 % of this facies. *Orbitolites* and *Gypsina* tests are up to 0.75 mm in length. All bioclasts are abraded. Micrite comprises 2 to 10 % of this facies, and is patchily distributed through samples. Clay-grade non-carbonate material comprises 7 to 25.5 % of this facies.

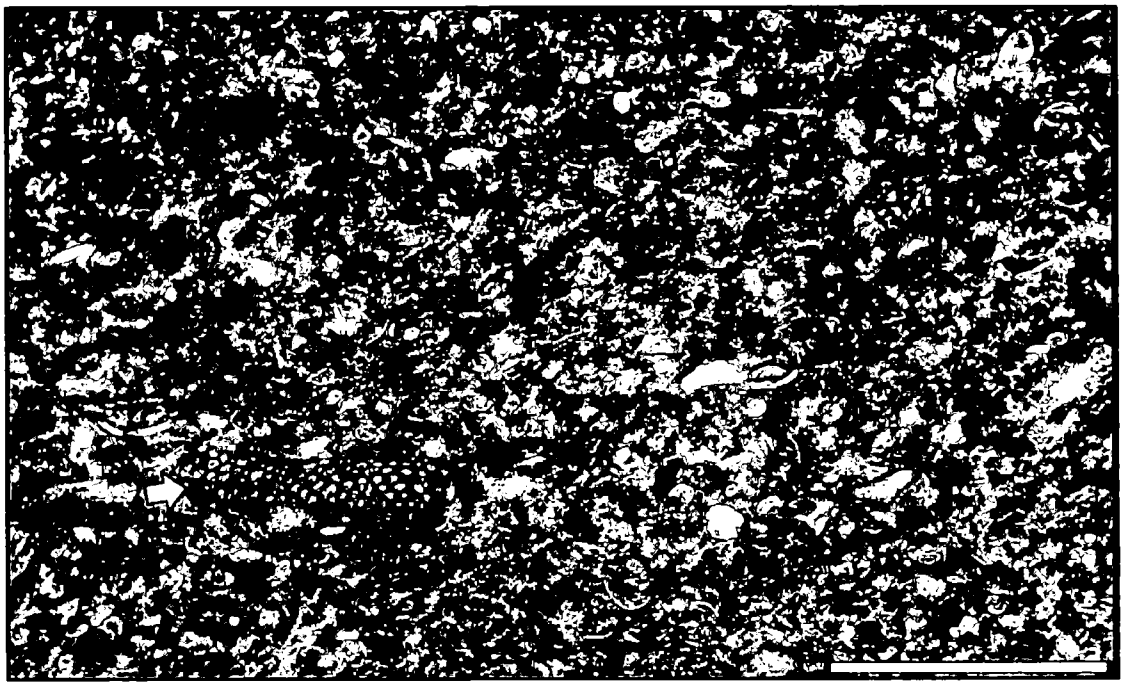


Figure 3.26a Low-magnitude photomicrograph (PPL) of the miliolid coralline algae pack/grainstone. This facies is characterized by an abundance of abraded miliolids and coralline algae with rare *Orbitolites* (arrowed) (sample LCA 36, log CA-3 bed 16). Scale bar=2 mm.

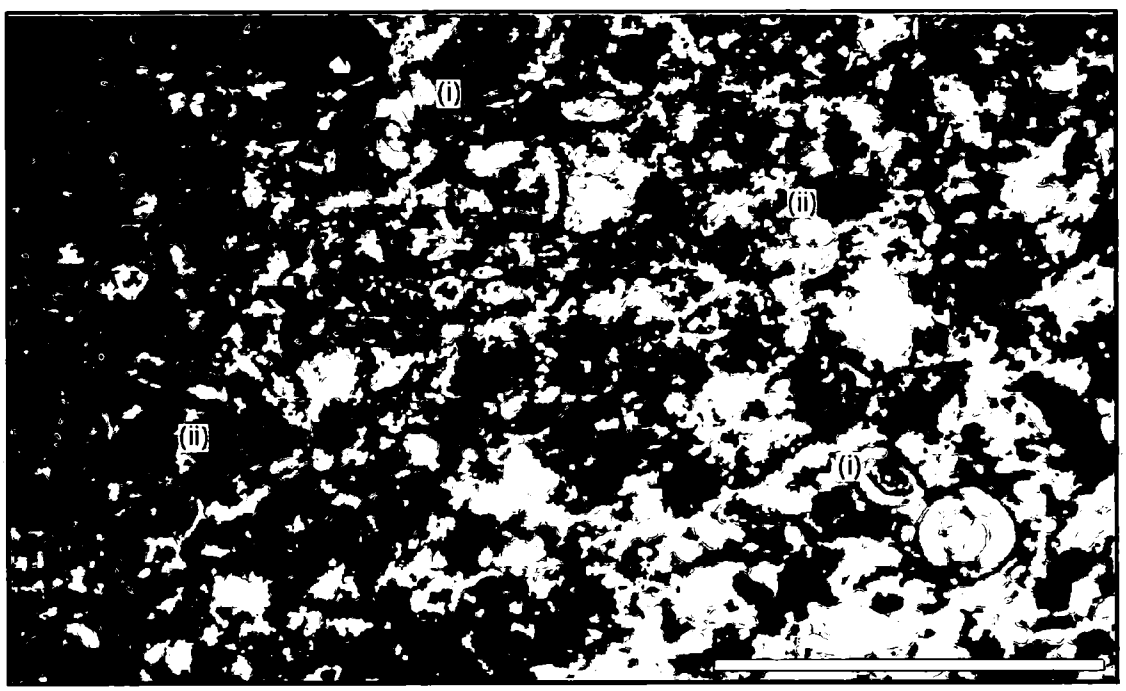


Figure 3.26b Photomicrograph (PPL) of the miliolid coralline algae pack/grainstone. Abraded miliolid tests (i) have slightly reddened tests. Coralline algae are always fragmented (ii) (sample LCA 36, log CA-3 bed 16). Scale bar=2 mm.

Diagenesis

Intergranular drusy calcite spar cement with localised equant spar cement comprises 35.5 to 50 % of this facies. Intragranular microspar cements are observed within foraminifera tests and echinoid spines demonstrate syntaxial cements. Originally aragonitic bioclasts have been leached with the resultant mouldic macropores completely occluded with clear, drusy calcite cement. Red-brown coloured clay-grade material occurs concentrated along mm-scale dissolution seams.

Interpretation: depositional environment

Alignment of nodules is interpreted as a remnant of bedding. The miliolid coralline algae pack/grainstone facies contains a marine faunal assemblage. Restricted marine circulation is inferred from the relatively scarcity of stenohaline biota, although the presence of texturaliids and *Gypsina* indicates a marine connection (Chaproniere 1975, Ghose 1977, Hallock and Glenn 1986, Murray 1991, Geel 2000). Deposition in a shallow marine setting with very high incident light is inferred from the abundance of miliolids and dearth of rotaliid foraminifera such as *Amphistegina* and *Nummulites* (Ghose 1977, Hallock and Glenn 1986, Geel 2000).

Moderate to high-energy, agitated conditions are inferred from the pack/grainstone texture and the fragmented nature of bioclasts, although the presence of clays (that would have been deposited from suspension) suggests (intermittent) low-energy periods. In addition, current energies were not sufficient to remove small foraminifera tests. Proximity to a reef-type environment is inferred from the presence of reef-derived debris (corals, large encrusting foraminifera). This sediment would have been too fine grained to supported large encrusting foraminifera such as victoriellids and *Chapmanina* that would have inhabited cryptic environments within coarse skeletal debris (Romero 2001, Romero *et al.* 2002).

In summary, the miliolid coralline algae pack/grainstone facies formed within a very shallow marine setting characterised by variable depositional energies, and received debris from nearby coral-rich areas. The comparable Skeletal Grainstone facies from the Nummulitique of the French Alps has been interpreted as a winnowed back shoal to lagoonal environment (Sayer 1995). Similar facies are also described from the Qum Formation, Iran, and are interpreted as inner-shelf shallow lagoon sediments (Okhravi and Amini 1998). However, true lagoon development in the

Calders area is thought unlikely due to the absence of topographic relief within the carbonate facies.

3.3.2 Mixed carbonate-siliciclastic larger benthic foraminifera facies group

Facies included in this grouping contain roughly equal amounts of carbonate and siliciclastic components, and have a sandy texture in outcrop. Siliciclastic material may constitute up to 50 % of facies within this group. Facies are characterised by the presence of larger benthic foraminifera, in particular *Nummulites, Discocyclina, Operculina* and *Gypsina*. Subordinate biota includes coralline algae, smaller benthic foraminifera, molluscs, echinoids and serpulids. The dominant textures observed are siliciclastic grainstone, packstone and wackestone (**Table 3.2**).

3.3.2.1 *Nummulites* siliciclastic pack/grainstone

Lithologies: *Nummulites* siliciclastic packstone
Nummulites siliciclastic wacke/packstone
Nummulites grainstone

Occurrence and bed characteristics

The *Nummulites* siliciclastic pack/grainstone facies is encountered in most logged sections in the Calders area, but is more prevalent in the lower half of the succession within carbonate intervals 1 to 4 (**Figure 3.9**). This facies occurs as massive, apparently tabular units that can be traced laterally for at least 2 km (across the extent of the studied area). Measured bed thickness ranges from 1 to 3.4 m. Lower bedding contacts with the lenticular polymict conglomerate facies are gradational over 10 to 15 cm. The contact may be enhanced by dissolution, with dark-brown clays concentrated along dissolution seams (**Figure 3.27a**). The upper contact with other larger foraminifera siliciclastic facies is gradational. Upper contacts with carbonate-dominated facies are sharp to gradational (over tens of centimetres), and may be delineated by dissolution seams along which clays are concentrated (**Figure 3.29c**). Crude cross-stratification is locally present (**Figure 3.27b**). A north to north-west directional dip direction (with dips up to 20°) is measured from cross sets.

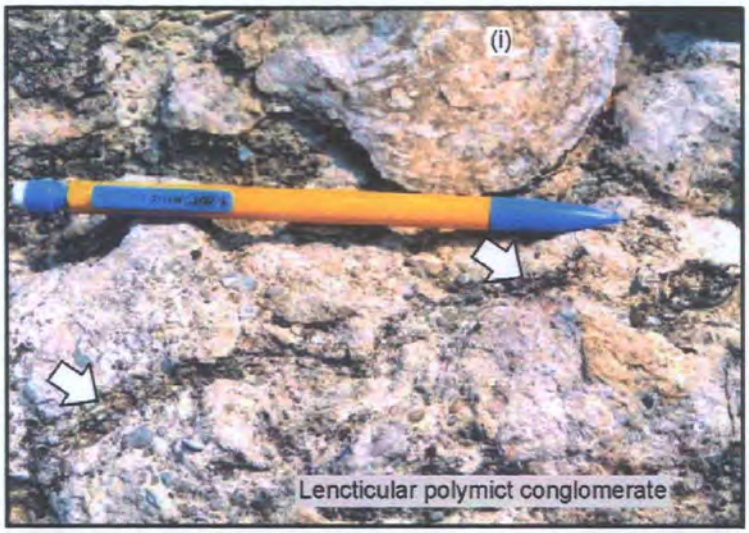


Figure 3.27a Outcrop of the contact between the lenticular polymict conglomerate and the overlying *Nummulites* siliciclastic pack/grainstone facies that has been enhanced by dissolution. Dark brown clays are concentrated along the dissolution seam (arrowed). Solitary corals (i) are present but rare in this facies. Photograph of bed 14, log CA-7. Pencil=14 cm.

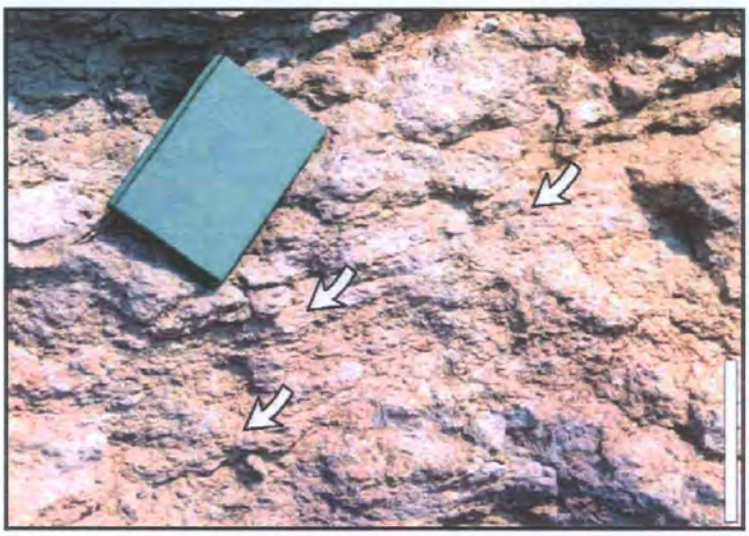


Figure 3.27b Planar cross-stratification of the *Nummulites* siliciclastic pack/grainstone facies (arrowed). Apparent progradation direction of fore sets is towards the north-northwest. Photograph of bed 21, log CA-7, facing east. Scale=15 cm.

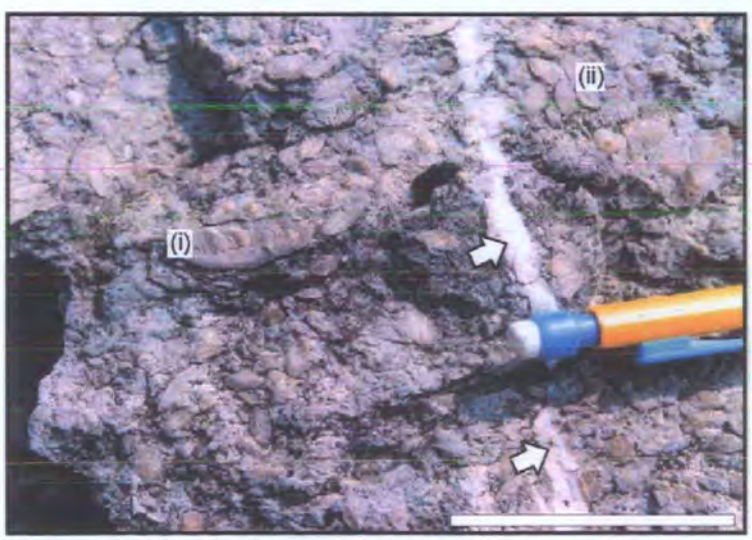


Figure 3.27c Detailed view of *Nummulites*. Note the presence of large lenticular to discoidal (i) and smaller robust (ii) forms. Bedding-normal fractures (arrowed) have been cemented with calcite. Photograph of bed 21, log CA-7. Scale=5 cm.

Lithological description

The exposed surface of the *Nummulites* siliciclastic pack/grainstone facies weathers to a reddish-brown/grey colour (**Figures 3.27b** and **c**). Fresh surfaces are a homogenous pale blue-grey colour. This facies is characterised by an abundance of *Nummulites* (17.5 to 45 % of grains) (**Figure 3.27c**). *Nummulites* tests have a robust to lenticular morphology and are up to 3 cm in diameter (**Figures 3.27c** and **3.29a**). The typical diameter is around 0.5 cm (**Figure 3.28a**). Larger *Nummulites* tests are more common towards the top of beds of this facies. Tests are locally imbricated and may be concentrated and aligned in large vertical burrows. Tests are well preserved, although fragments are present in the matrix. Fragments (up to coarse sand grade) are angular to slightly abraded (**Figure 3.29a**).

The *Nummulites* siliciclastic pack/grainstone facies contains a diverse benthic foraminifera assemblage that includes *Gypsina* (0 to 10.5 %), *Discocyclina* (0 to 6.5 %), *Operculina* (<1 %), *Amphistegina* (< 1 %), *Calcarina* (0 to 5.5 %), miliolids (0 to 5 %), texturaliids (0 to 2 %) and undifferentiated small benthics (< 1 %). *Gypsina* occurs as laminar forms (**Figure 3.28b**). *Discocyclina* and *Operculina* tests have a flattened lenticular morphology, and are often fragmented. Angular fragments are up to 1 mm in length. Miliolids and *Gypsina* have an abraded appearance. Large encrusting foraminifera (*Fabiania* and *Haddonia*) are often constituents of foralgal rhodoliths. Spheroidal rhodoliths are rare (<1 % of grains), and are composed of tight, concentric alternating algal and foraminifera laminae.

Other biota present in the *Nummulites* siliciclastic pack/grainstone facies includes coralline algae (0 to 10 %), echinoids (0 to 5.5 %), fenestrate bryozoa (0 to 7.5 %) and molluscs (0 to 3.5 %). Large robust oyster valves are observed *in situ* within beds of this facies (**Figure 3.29b**). Coralline algae occur as rounded warty-branching fragments that have good preservation of internal structure. Coarse sand-grade coralline algae peloids (0 to 7.5 %) are very well sorted. Spheroidal to elliptical structures composed of fabric destructive dolomite are interpreted as coprolites (**Figure 3.28b**).

Micrite (0.5 to 44.5 %) is locally absent (**Figure 3.28a** and **b**). Micrite is grey to dark brown in thin section (**Figures 3.28c**). Non-carbonate material comprises 14 to 50 wt. % of this facies. Siliciclastic grains are clay to coarse sand-grade, and comprise mono-and polycrystalline quartz, orthoclase feldspar and lithics (siltstone

Figure 3.28a Photomicrograph (PPL) of the *Nummulites* siliciclastic pack/grain-stone facies. *Nummulites* tests (i) are typically 2-5 mm in diameter, and occur in association with laminar *Gypsina* (ii) and *Rotorbinella* (iii) (sample LCA 30, log CA-3 bed 7). Scale bar=2 mm

Figure 3.28b Photomicrograph (PPL) of the *Nummulites* siliciclastic pack/grain-stone facies. Dolomitised coprolites (i) and rounded *Gypsina* tests (ii). This facies contains abundant fine to medium sand-grade material (iii) (sample LCA 30, log CA-3 bed 7). Scale bar=2 mm

Figure 3.28c (PPL) Dissolutional grain contacts (arrowed). Insoluble clays are concentrated along the dissolution seams (sample LCA 230, log CA-15) Scale bar=2 mm

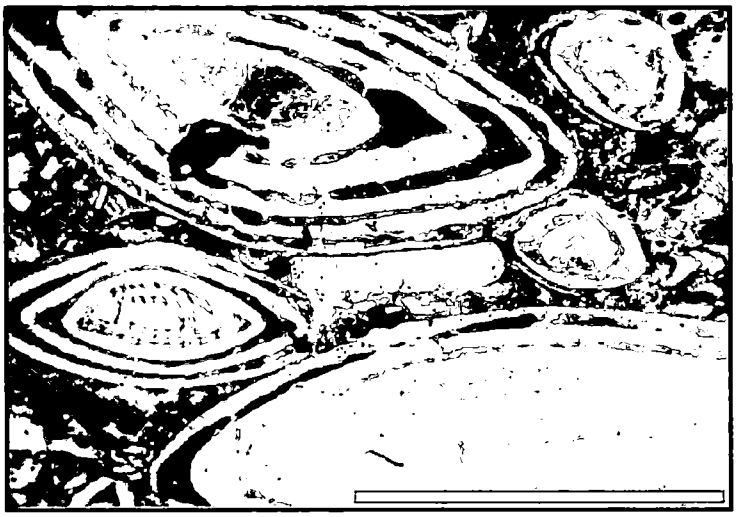


Figure 3.29a Photomicrograph (PPL) of the *Nummulites* siliciclastic pack/grain-stone facies illustrating the variability of *Nummulites* test size. The matrix (grey-brown colour) contains abundant foraminifera debris (sample LCA 22, log CA-2 bed 7). Scale bar=2 mm

Figure 3.29b Large, robust oyster within the *Nummulites* siliciclastic pack/grainstone facies (bed 14, log CA-7). Scale bar=5 cm

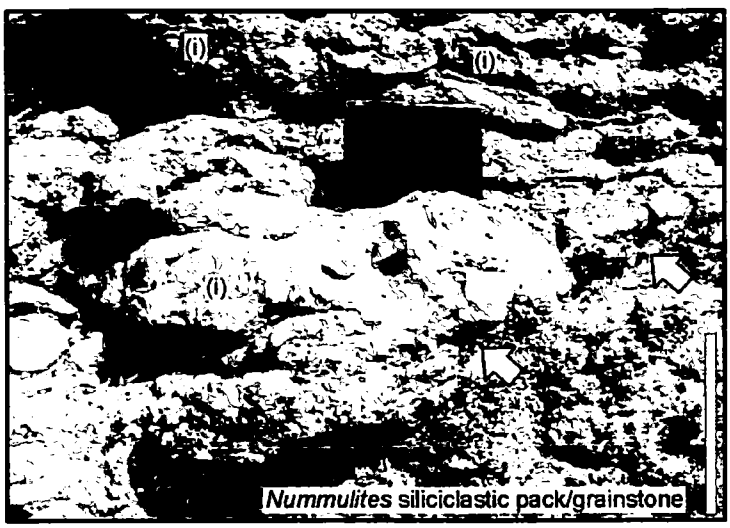


Figure 3.29c Field exposure of the upper contact of the *Nummulites* siliciclastic pack/grainstone and coral wacke/floatstone facies (log CA-7, beds 13 and 14). The contact (arrowed) is enhanced by dissolution with dark siliciclastic grains concentrated along the dissolution seam (arrowed). The initial coral colonizers (i) have platy morphologies. Scale bar=15 cm.

and quartz-mica schist). The micas have commonly been altered to rusty-brown coloured clay observed in thin section. Siliciclastic grains are angular, moderate to well sorted and have a low sphericity.

Diagenesis

Intragranular microspar cements are observed within foraminifera tests and echinoid spines demonstrate syntaxial calcite overgrowth cements. The contacts between larger foraminifera tests may be sutured along dissolution seams along which clays are concentrated (**Figures 3.28c**). Winnowed intergranular areas have been cemented with equant calcite spar, and patchy recrystallisation of the micrite matrix is evident.

Interpretation: depositional environment

The *Nummulites* siliciclastic pack/grainstone facies contains a marine faunal assemblage. Normal open marine conditions are inferred from the diversity of stenohaline biota. Deposition within the photic zone is inferred from the abundance of symbiont bearing larger benthic foraminifera (*Nummulites, Discocyclina, Operculina, Amphistegina* and *Calcarina*) (Ghose 1977, Hallock and Glenn 1985, 1986; Murray 1991, Geel 2000). The flattened test morphology of *Discocyclina* and *Operculina* are characteristic of deeper parts of the photic zone, towards the lower parts of their habitat range (Hallock and Glenn 1985, 1986). The intermediate, robust test morphology of *Nummulites* is characteristic of moderately agitated conditions within intermediate depths of their habitat range (Hottinger 1983, Hallock and Glenn 1985). The robust morphology of large *Nummulites* (within the 1 to 3.5 cm range) is suggestive of shallow waters with stressed environmental conditions, delaying reproduction (Hallock 1988, Hallock and Glenn 1986, Purton and Brasier 1999). Environmental stress may relate to a combination of limited nutrients, substrate instability and turbidity (Hallock 1988).

Proximity to a siliciclastic source is concluded from the significant proportion of siliciclastic material in this facies. It is suggested that fine-grained siliciclastic material was deposited from suspension during relatively quiet periods. Agitated conditions are confirmed from the presence of concentric, laminar rhodoliths and warty coralline algae (Minnery *et al.* 1985, Minnery 1990).

It is postulated that substrate instability, as a consequence of siliciclastic input and a moderate to high-energy hydrodynamic regime, inhibited the development of a sessile calcareous benthic community. *Nummulites,* inferred to have undergone limited transport and reworking prior to deposition and lithification, is thought to have been able to extract itself if buried (see **Chapter 6** for discussion). Miliolids, *Amphistegina*, *Gypsina* and *Calcarina* can adapt to an epiphytic mode of life (Ghose 1977, Reiss and Hottinger 1984, Kitazato 1988, Hohenegger *et al.* 1999). It is postulated that these foraminifera were able to inhabit an environment with unstable substrates through attaching themselves to sea floor vegetation. These foraminifera have undergone more abrasion than *Nummulites*, thus are not preserved *in situ.*

In summary, the *Nummulites* siliciclastic pack/grainstone facies formed within a moderate to high-energy marine environment with a high siliciclastic input and an unstable substrate. This facies has similarities to the Nummulite Packstone facies of Sayer (1995), which is interpreted as part of an authochthonous Nummulite-ramp system. The *Nummulites* siliciclastic packstone grainstone facies contains a significantly higher proportion of coarse clastic material than the Nummulite Packstone facies. Similar nummulitic facies, interpreted as reworked nummulitic bank deposits, are described from the Seeb Limestone of Oman (Racey 1988, 2001).

3.3.2.2 *Nummulites Discocyclina* siliciclastic packstone

Lithologies: *Nummulites Discocyclina* siliciclastic packstone
Nummulites Discocyclina siliciclastic grainstone

Occurrence and bed characteristics

The *Nummulites Discocyclina* siliciclastic packstone facies is differentiated from the siliciclastic *Nummulites* pack/grainstone facies by a relative abundance of orthophragminids. This facies is particularly common in the lower portions of the Calders studied succession within carbonate intervals 1 to 3 (**Figure 3.9**). Measured bed thickness is consistently around 1.7 m and upper and lower bedding contacts are transitional. This facies can be traced laterally up to 500 m. The *Nummulites Discocyclina* siliciclastic packstone typically succeeds the siliciclastic *Nummulites* packstone grainstone facies.

Lithological Description

The exposed surface of the *Nummulites Discocyclina* siliciclastic packstone facies weathers to a buff/light brown red colour. Fresh surfaces are a light blue-grey colour. This facies is characterised by an abundance of *Nummulites* (6 to 10 %) and *Discocyclina* (3.5 to 25 %). *Nummulites* are robust/lenticular forms up to 3 mm in diameter (**Figures 3.30a** and **b**). *Nummulites* are always fragmented. Fragments are abraded and poorly sorted. *Discocyclina* tests are up to 2 cm in diameter. Tests demonstrate thin, discoidal morphologies (**Figure 3.30b**), and are typically preserved intact. Larger foraminifera tests demonstrate chaotic stacking (**Figure 3.30b**).

Subordinate bioclasts present in this facies are miliolids (0 to 2.5 %), texturaliids (< 0.5 %), *Amphistegina* (< 1 %), *Operculina* (0 to 2 %), *Gypsina* (0 to 2 %), echinoids (0 to 2 %), molluscs (0 to 3.5 %), corals (0 to 2.5 %), bryozoa (0 to 2.5 %) and small hyaline benthic foraminifera (< 0.5 %). Miliolids and *Amphistegina* tests are preserved intact but are abraded. *Operculina* is always fragmented. Fragments are < 1 mm in diameter (**Figure 3.30b**).

Micrite comprises up to 58 % of this facies (**Figures 3.30a** and **b**). The total non-carbonate content of this facies ranges from 14 to 38.5 wt. %. Siliciclastic grains are clay to silt grade.

Diagenesis

Intragranular microspar cement is observed within foraminifera tests. Originally aragonitic bioclasts have been leached with the resultant mouldic pores completely occluded with drusy calcite cement. Echinoid spines may have syntaxial calcite overgrowths. In some samples the originally micritic matrix has recrystallised into microspar (up to 64.5 %).

Interpretation: depositional environment

The *Nummulites Discocyclina* siliciclastic packstone facies contains a marine faunal assemblage. Normal marine conditions are inferred from the abundance of rotaliid foraminifera and the diversity of stenohaline biota.

Deposition within the photic zone is inferred from the presence of symbiont-bearing larger formanifera (*Nummulites, Discocyclina* and *Amphistegina*). Moderate to deeper-water depths and/or low incident light are inferred from the dominance of

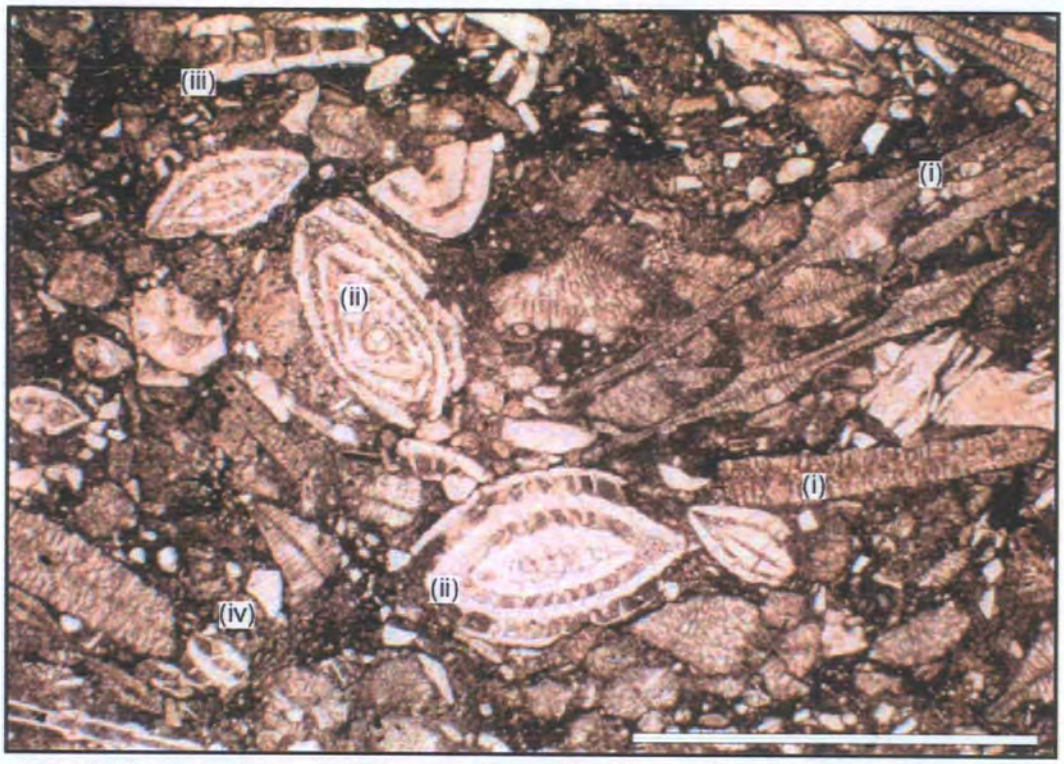


Figure 3.30a Photomicrograph (PPL) of the *Nummulites Discocyclina* siliciclastic packstone facies. Large, flattened *Discocyclina* (i) occurs with lenticular *Nummulites* (ii) and fragmented *Operculina* (iii). Siliciclastic grains are angular (iv) (sample LCA 235, log 15 bed 1). Scale bar=2 mm.

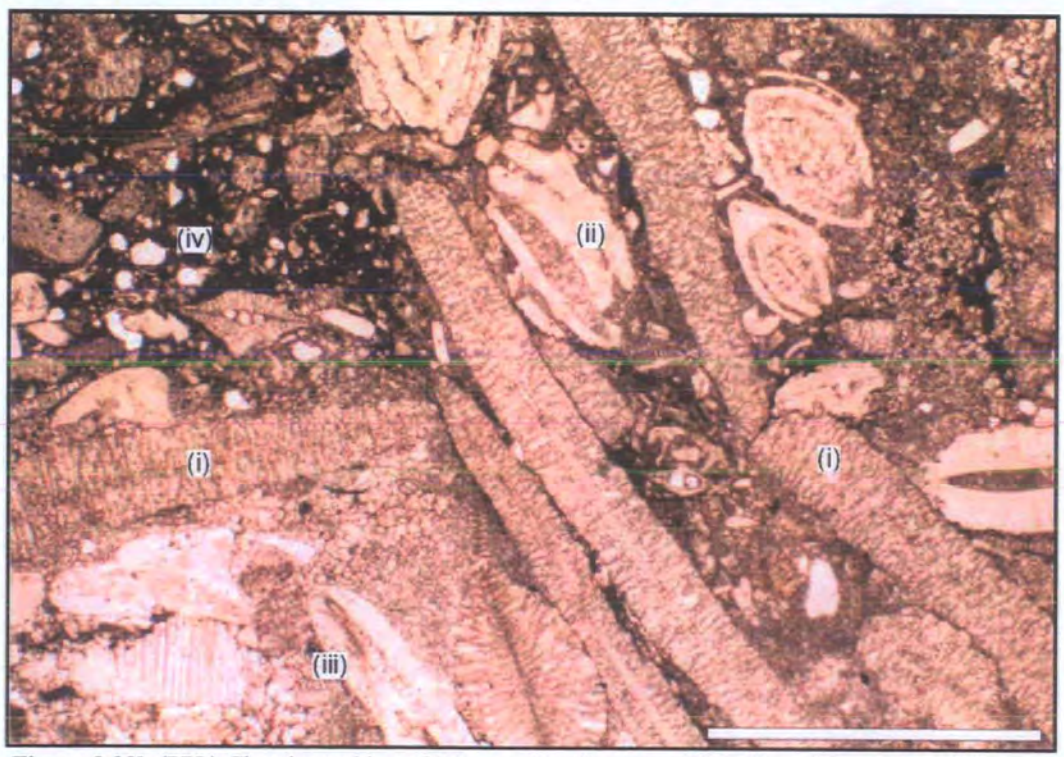


Figure 3.30b (PPL) Chaotic stacking of *Discocyclina* (i), *Nummulites* (ii) and *Operculina* (iii) tests. Note the presence of dark-brown clays concentrated in the matrix (iv) (sample LCA 235, log 15 bed 1). Scale bar=2 mm.

flattened and intermediate forms of *Discocyclina* over relatively small robust forms of *Amphistegina* and *Nummulites*. In general, orthophragminid foraminifera such as *Discocyclina* inhabited lower levels of the photic zone within mid- to outer shelf environments (Aigner 1982, Ghose 1977, Hallock and Glenn 1986, Racey 1988, Bryan 2001). This is consistent with the paucity and abrasion of miliolids, which are more common in shallower water/inner shelf environments with high incident light (Ghose 1977, Hallock and Glenn 1985). It is suggested that miliolids and the more spherical forms of *Nummulites* and *Amphistegina* and coral fragments have been transported into this environment from a high-energy, shallow part of the photic zone.

Low-energy conditions are confirmed from the significant percentage of clay to silt-grade siliciclastic material that would have been deposited from suspension. It is postulated that turbidity would have been high in this environment as a consequence of this fine-grained input. Low amplitude *in situ* reworking of foraminifera tests is interpreted from the localised stacking of tests (**Figure 3.30b**) Chaotic stacking of large discoidal *Discocyclina* tests has been interpreted as low-amplitude storm reworking within Eocene deposits of the Nummulitique (Sayer 1995), Egypt (1983) and Oman (Racey 1988, 2001).

In summary, the *Nummulites Discocyclina* siliciclastic packstone facies formed within an open marine, moderate to low-energy environmental setting within the intermediate to deeper parts of the photic zone but quite close to the shoreline. Similar facies are described from the Seeb limestone of Oman (Racey 1988, 2001) and the Nummulitique of the French Alps (Sayer 1995). These facies are interpreted as low-relief foraminiferal banks deposited within a mid-shelf shelf below fair weather wave base.

3.3.2.3 *Operculina* mudstone

Lithologies: *Operculina* mudstone
Operculina Nummulites wackestone

Occurrence and bed characteristics

This facies is difficult to identify and sample as it weathers recessively due its muddy texture. This facies is identified in the upper portions of the Calders section at the base of carbonate intervals 5 and 6 (**Figure 3.9**). This facies succeeds a siliciclastic facies. Measured bed thickness ranges from 0.1 to 1.6 m. Beds have a

muddy appearance in outcrop. Upper and lower bedding contacts are sharp and non-erosive.

Lithological description

The exposed surface of the *Nummulites Operculina* siliciclastic wackestone facies weathers to a light-brown/buff colour. Fresh surfaces are a buff colour with dark grey patches. This facies is largely unfossiliferous but contains rare large discoidal *Operculina* (up to 5 %) and *Nummulites* (< 1 %). *Operculina* tests have a flattened lenticular morphology and are up to 7 mm in diameter (**Figure 3.31a**). *Operculina* tests typically occur intact, but minor fragments up to 1 mm in length are observed in the matrix.

Other bioclasts present in this facies are miliolids (< 1 %), *Gypsina* (< 1 %), laminar coralline algae (< 0.5 %), echinoid spines (<1 %), corals (< 1 %), brachiopod spines (< 0.5 %), bryozoa (0.5 %) and bivalves (1 %). Rare delicate bivalves are preserved articulated (**Figure 3.31b**). Most fossils are fragmented and abraded.

Dark brown mixed carbonate-siliciclastic mud comprises up to 85 to 90 % of this facies (**Figures 3.31a** and **b**). Rare (< 5 %) angular silt grade siliciclastic grains are evident in thin section (**Figure 3.31b**). Grains are monocrystalline quartz (3 %) and lithic grains (1 %) with relatively minor orthoclase feldspar (1 %). The total non-carbonate content of this facies ranges from 37 to 37.5 wt. %.

Diagenesis

The matrix of the *Nummulites Operculina* siliciclastic wackestone facies includes localised patches of neomorphic microspar. Intragranular calcite cements are observed within foraminifera tests and echinoid spines occasionally have syntaxial calcite overgrowths.

Interpretation: depositional environment

The *Operculina* mudstone facies comprises a marine faunal assemblage. Normal open marine conditions are interpreted from the presence of *Operculina* and *Nummulites* (Chaproniere 1975, Murray 1991, Hohenegger *et al.* 1999, Geel 2000), and the diversity of stenohaline biota. Deposition within the photic zone is inferred from the presence of symbiont-bearing foraminifera. Large, flattened discoidal

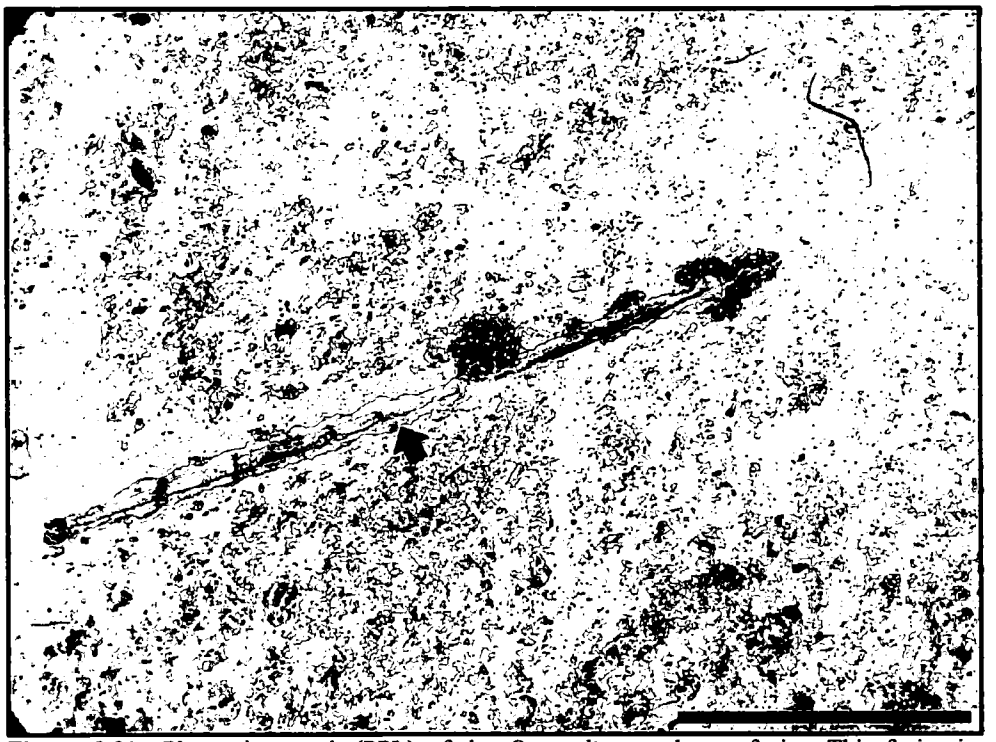


Figure 3.31a Photomicrograph (PPL) of the *Operculina* mudstone facies. This facies is characterized by the presence of large discoidal *Operculina* (arrowed) (sample LCA 6, log CA-6 bed 3). Scale bar=2 mm.

Figure 3.31b Photomicrograph (PPL) of a delicate bivalve preserved intact and articulated (arrowed) within the *Operculina* mudstone facies. (sample LCA 7, log CA-6 bed 5). Scale bar=2 mm.

Operculina tests are characteristic of the deeper parts of their habitat range, in the lower part of the photic zone (Hottinger 1983, Hallock and Glenn 1985, 1986, Geel 2000). Low-energy conditions are inferred from the high percentage of fine-grained material in this facies (which is interpreted to have been deposited from suspension) and the preservation of articulated delicate bivalves. *Operculina* typically has a preference for soft substrates in low-energy environments (Chaproniere 1975, Murray 1991, Hohenegger 1999).

It is inferred from the lack of abrasion of *Operculina* tests that they are *in situ*. *Nummulites* tests in comparison are slightly abraded and commonly fragmented, and therefore are interpreted to have been transported down-slope from a high-energy to a lower-energy environment. Most bioclasts are fragmented and abraded, thus it is inferred that they were also transported from a high to low-energy environment.

In summary, the *Operculina* mudstone facies is interpreted as a marine, low-energy moderately deep-water environmental deposit, probably from below fair weather wave base, in the lower part of the photic zone. *Operculina* is more-or-less *in situ* but abraded bioclasts, including *Nummulites*, have been transported downslope from a relatively shallow environment, accompanied by minor reef detritus. This facies has affinities with the Larger Foraminifera Wackestone facies of Hallock and Glenn (1985), interpreted as slope deposits and the *Discocyclina Operculina* Wackestone facies of Sayer (1995), interpreted as a mid- to outer ramp, deep water facies.

3.3.2.4 *Nummulites* coralline algae siliciclastic packstone

Lithologies: *Nummulites* coralline algae siliciclastic packstone
Nummulites coralline algae siliciclastic wacke/packstone
Nummulites rhodolithic siliciclastic pack/grainstone

Occurrence and bed characteristics

The *Nummulites* coralline algae siliciclastic packstone facies is common and is present in carbonate intervals 1 to 4. This facies is always associated with the *Nummulites* siliciclastic packs/grainstone. Measured bed thickness ranges from 0.85 to 1 m. Beds are largely massive but may have a nodular weathered appearance. Lower bedding contacts are gradational over 5 to 10 cm. Upper bedding contacts are sharp and non-erosional.

Lithological description

Exposed surfaces of the *Nummulites* coralline algae siliciclastic packstone facies weather to a pale brown-grey colour. Fresh surfaces are a blue-grey colour with localised pink-grey patches. This facies is characterised by an abundance of *Nummulites* (11 to 40 %) and coralline algae (5 to 25 %) (**Figures 3.32a** and **b**). *Nummulites* tests up to 4 mm in diameter have a robust lenticular morphology. In axial cross-section some *Nummulites* tests appear to have an enlarged (megalospheric) proloculus (**figure 3.32b**). *Nummulites* tests occur intact and fragmented. Fragments up to 1 mm are angular to abraded.

Coralline algae occur as fragmented thick, warty branches up to 6 mm (**Figure 3.32a**), thin laminar crusts (**Figure 3.32b**), and small foralgal rhodoliths (**Figure 3.33**). Thick branches demonstrate well-preserved internal structures and may be bored by lithophagid bivalves (**Figure 3.32a**). Thin laminar coralline algae fragments are up to 0.1 mm thick are abraded poorly preserved primary structures (**Figure 3.32b**). Subspherical to elliptical rhodoliths with delicate, warty branches are up to 5 cm in diameter. Rhodoliths may contain *Gypsina* and/or *Fabiania* (**Figure 3.33**). Incorporation of foraminifera into rhodoliths has modified the morphology, with stout branches developing on initially tight, concentric forms (**Figure 3.33c**). Well-sorted, coarse sand-grade coralline algae peloids (0 to 7.5 %) are present in this facies.

The *Nummulites* coralline algae siliciclastic packstone facies contains a diverse benthic foraminifera assemblage that includes texturaliids (0 to 1 %), *Gypsina* (0 to 13.5 %), *Discocyclina* (0 to 5 %), *Amphistegina* (0 to 2 %), *Calcarina* (0 to 5 %) and miliolids (0 to 1.5 %). Encrusting foraminifera (*Fabiania* and Haddonia) comprise up to 5 % of this facies. *Discocyclina, Amphistegina* and *Calcarina* are always fragmented. Fragments up to 1 mm in length are angular to abraded. *Discocyclina* tests demonstrate robust, lenticular morphologies. Miliolids are small (< 0.25 mm), typically fragmented and abraded. Subordinate bioclasts present in this facies are echinoids (0 to 1.5 %), molluscs (0 to 1.5 %), brachiopod spines (< 0.5 %) and bryozoa (0 to 3.5 %). In general, bioclasts are poorly sorted and show no preferred orientation.

Micrite comprises up to 40 % of this facies. The total non-carbonate material in this facies ranges from 15 to 26.5 wt. %. Siliciclastic grainsize ranges from clay to

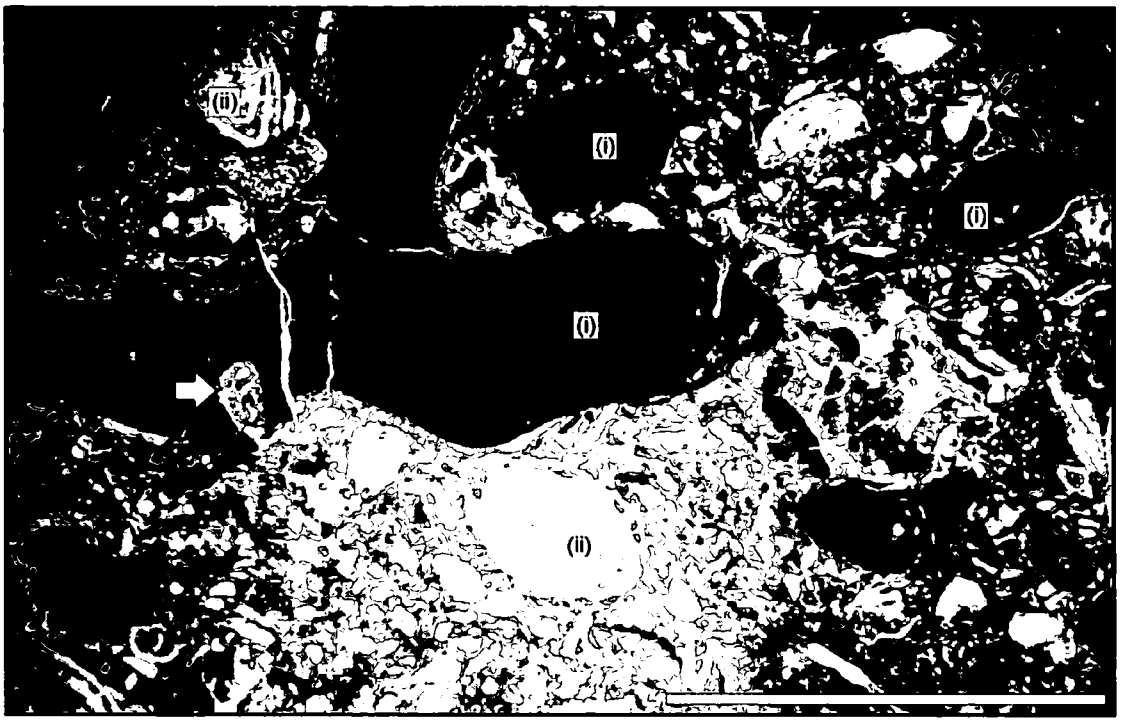


Figure 3.32a Photomicrograph (PPL) of the *Nummulites* coralline algae siliciclastic packstone facies. This facies is characterized by the presence of thick, nodular coralline algae growths (i) and *Nummulites* (ii). Coralline algae are often bored (arrowed). In this instance, the almond-shaped boring suggests a lithophagid bivalve (sample LCA 31, log CA-3 bed 7). Scale bar=5 mm.

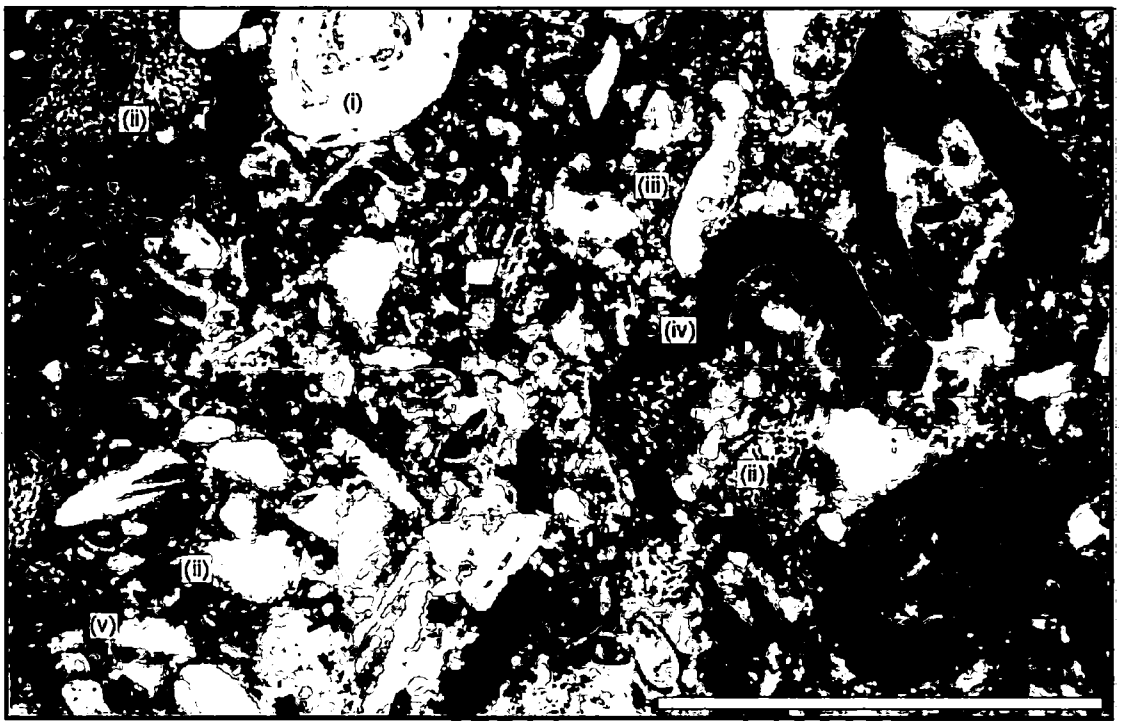


Figure 3.32b Photomicrograph (PPL) of the *Nummulites* coralline algae siliciclastic packstone facies. *Nummulites* are often the megalospheric form (i), and occur in association with *Gypsina* (ii) and *Haddonia* (iii). Coralline algae are also present as laminar crusts (iv). Sample LCA 31, log CA-3 bed 7). Scale bar=5 mm.

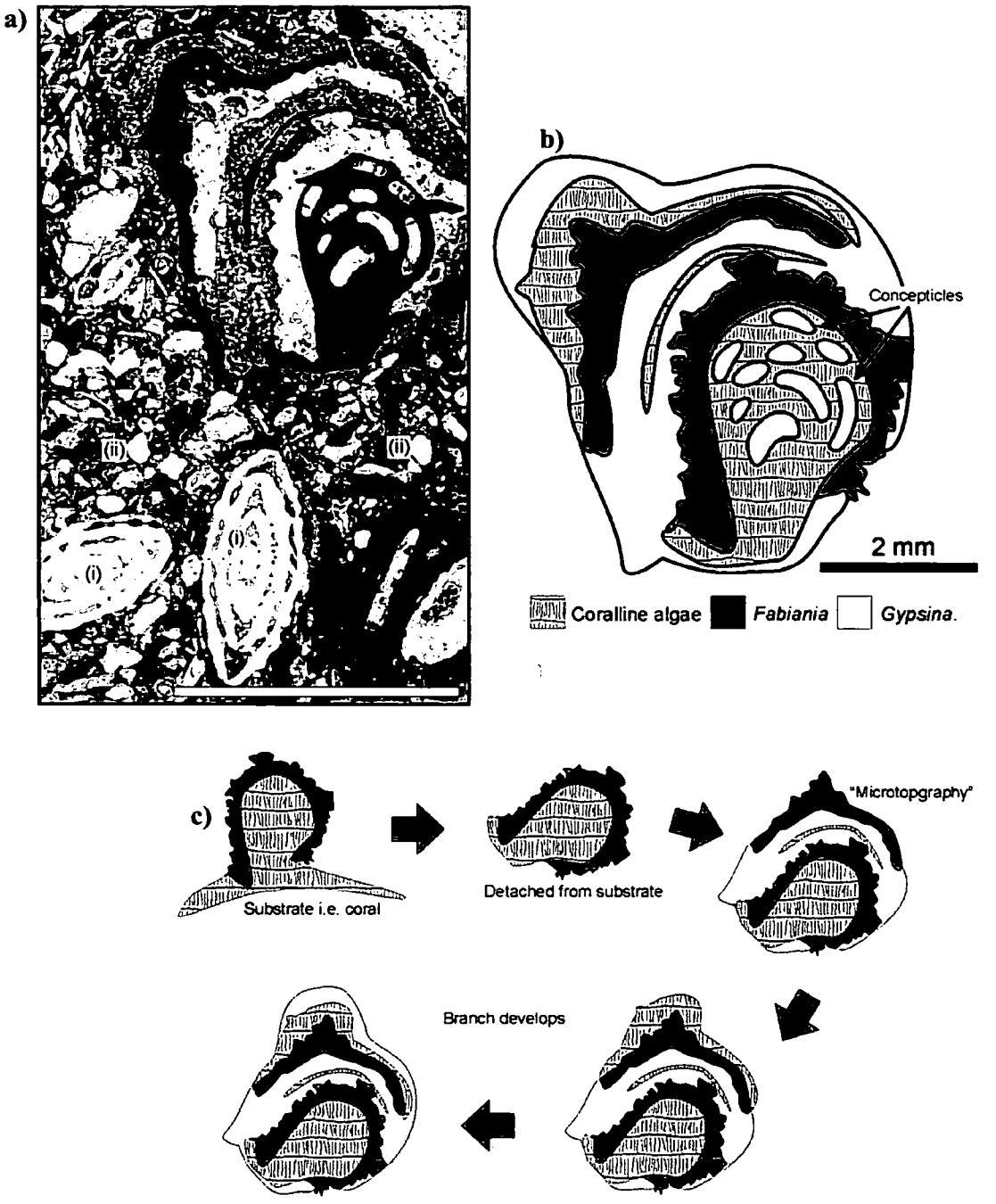


Figure 3.33a Photomicrograph of the *Nummulites* coralline algae siliciclastic packstone facies. Microspheric, lenticular *Nummulites* (i) are present in addition to megalospheric forms. Siliciclastic grains (ii) are angular and well sorted. Sample LCA 88, log CA-7 bed 13 (top)). Scale bar=5 mm.

Figure 3.33b Interpretation of the foralgal rhodolith in a).

Figure 3.33c Stylized 'encrustation history' of the foralgal rhodolith. The initial warty coralline algae branch was attached to a hard substrate, possibly a coral and became detached. The growth of encrusting foraminfera (*Fabiania* and *Gypsina*) produced a small area of topography that provided a relatively stable site for further crust development. These successive crusts developed into a stout branch.

medium sand. Sand-grade siliciclastic grains identified are predominantly mono- and polycrystalline quartz with lithics and orthoclase feldspar. Lithic grains are siltstone and quartz-mica schist. The micas have commonly weathered to a yellow to rusty-brown coloured clay observed in thin section. Siliciclastic grains are angular, moderately well sorted and have a low sphericity (**Figure 3.33a**).

Diagenesis

Intragranular microspar is observed within foraminifera tests, and echinoid spines commonly have well-developed syntaxial calcite overgrowths. Intergranular areas are locally composed of neomorphic microspar (0 to 20.5 %).

Interpretation: depositional environment

The *Nummulites* coralline algae siliciclastic packstone facies contains a marine faunal assemblage. Normal marine conditions are inferred from the abundance of rotaline foraminifera (*Nummulites, Amphistegina* and *Calcarina*), and the diversity of stenohaline biota (Murray 1991). Deposition within the photic zone is inferred from the presence of symbiont-bearing larger foraminifera and coralline algae (Chaproniere 1975, Hottinger 1983, Hallock and Glenn 1985, 1986). The robust lenticular morphology of *Nummulites* is indicative of moderate energy and intermediate to shallow depths within their habitat range (Hottinger 1983, Hallock and Glenn 1985, 1986). It is suggested from the limited abrasion of *Nummulites,* and the well-preserved nature of warty *Mesophyllum* branches, that they have undergone limited transport and reworking prior to lithification

An agitated marine environment is deduced from abraded bioclasts, foralgal rhodoliths and thick, warty coralline algae branches (Bosence 1983, 1985, Reid and MacIntyre 1988, Minnery *et al.* 1985, Minnery 1990). It is suggested that thin, laminar coralline algae fragments and larger encrusting foraminifera floating in the matrix (i.e. *Fabiania, Haddonia* and *Gypsina*) are derived from fragmented rhodoliths (Minnery *et al.* 1985, Minnery 1990).

A siliciclastic input is inferred form the significant non-carbonate component of this facies. It is postulated that unstable substrates, related to constant input of siliciclastics and the moderate-energy hydrodynamic regime, inhibited the colonisation of this environment by large sessile calcareous benthic organisms such as corals. The presence of sea grass in this facies is inferred from the modest abundance

of *Amphistegina, Calcarina* and miliolids (Chaproniere 1975, Murray 1991, Hohenegger *et al.* 1999, Geel 2000). These foraminifera are preserved fragmented and abraded and have undergone reworking prior to lithification. Sea grasses may also have provided a substrate for coralline algae (Carbone *et al.* 1994), and may have acted as a baffle, trapping suspended sediment.

In summary, the *Nummulites* coralline algae siliciclastic packstone facies formed within an agitated, shallow marine environment within the shallow to intermediate depths of the photic zone, and with a high siliciclastic input and unstable substrates. This facies has similarities with the Nummulite-algal Packstone facies of Sayer (1995), which is interpreted as debris locally derived from a shallow shoal area.

3.3.2.5 Coralgal foraminifera siliciclastic float/rudstone

Lithologies: Coralgal foraminifera siliciclastic float/rudstone

Occurrence and bed characteristics

The coralgal foraminifera siliciclastic float/rudstone facies is identified in carbonate interval 5 (**Figure 3.9**). This facies is a massive unit with a local nodular weathered texture. Nodules are up to 10 cm in length and may contain a coral colony. Alignment of nodules is a remnant of bedding. Measured bed thickness is consistently around 1.5 m. Upper and lower bedding contacts are sharp and non-erosional.

Lithological description

The exposed surface of this facies weathers to a light brown colour. Fresh surfaces are a buff colour. The coralgal foraminifera siliciclastic float/rudstone facies is characterised by sparsely distributed, *in situ* branching and foliaceous coral colonies situated within a siliciclastic float/rudstone matrix. Porites and Actinacis colonies up to 15 cm in height dominate corals. The total coral skeletal volume ranges from 10 to 15 %, and corals do not make up a framework.

The bioclastic float/rudstone matrix contains abundant coral (18.5 to 28.5 %), coralline algae (2 to 5%) and foraminiferal debris (**Figure 3.34a**). Abraded coral fragments up to 1 cm form the nuclei for foralgal rhodoliths (**Figure 3.34b**). Tightly concentric, elliptical rhodoliths range from 0.5 to 5 cm in diameter. Rhodolith laminae

up to 3 mm thick are composed of a combination of coralline algae, *Haddonia* and acervulinids (**Figure 3.34**). The incorporation of foraminifera into the rhodoliths has led to the development of stocky branches. Abraded thin, laminar coralline algae fragments and medium to coarse grade coralline algae peloids (21 to 28.5 %) are present.

Foraminifera present in this facies are miliolids (0.5 %), texturaliids (< 0.5 %), *Gypsina* (0 to 2 %), *Nummulites* (2 %), *Amphistegina* (< 0.5 %) and *Calcarina* (< 0.5 %). Large encrusting foraminfera (victoriellids, *Haddonia* and *Fabiania*) comprise 1.5 to 3.5 % of this facies. Miliolids are abraded with reddened tests. Lenticular, slightly flattened *Nummulites* tests up to 3 mm and *Amphistegina* tests up to 0.5 mm are preserved intact but slightly abraded. *Calcarina* tests are up to around 2 mm. Subordinate biota present in this facies are echinoids (0.5 to 2 %), brachiopod spines (< 0.5 %), *Cerithium* gastropods (1 %), bivalves (< 0.5 %), bryozoa (0.5 %) and sponge spicules (< 0.5%).

Micrite comprises 2 to 7 % of this facies. The total siliciclastic content of this facies ranges from 22 to 23.5 wt. %. Siliciclastic grainsize ranges from clay (up to 9 %) to coarse sand (up to 14.5 %). Siliciclastic grains are mono- and polycrystalline quartz (< 8.5 %), lithics (< 6 %) and minor orthoclase feldspar (< 0.5 %). Lithic grains are micaceous siltstones. The micas have commonly weathered to rusty brown coloured clay and orthoclase feldspar grains have a dusky, altered appearance. Sand-grade siliciclastic grains are angular, moderately well sorted with low sphericity (**Figures 3.34a** and **b**).

Diagenesis

Coral and molluscan fragments have been dissolved out and the resultant mouldic pores have been occluded with drusy calcite cement (**Figure 3.34a**). Intragranular microspar cements occur within foraminifera tests and echinoid spines have syntaxial calcite cements. Red-brown clays occur concentrated along dissolution seams. The originally micritic matrix has locally recrystallised into microspar (**Figure 3.34a**).

Interpretation: depositional environment

The coralgal foraminifera siliciclastic float/rudstone facies contains a marine faunal assemblage. Normal open marine conditions are inferred from the diversity of

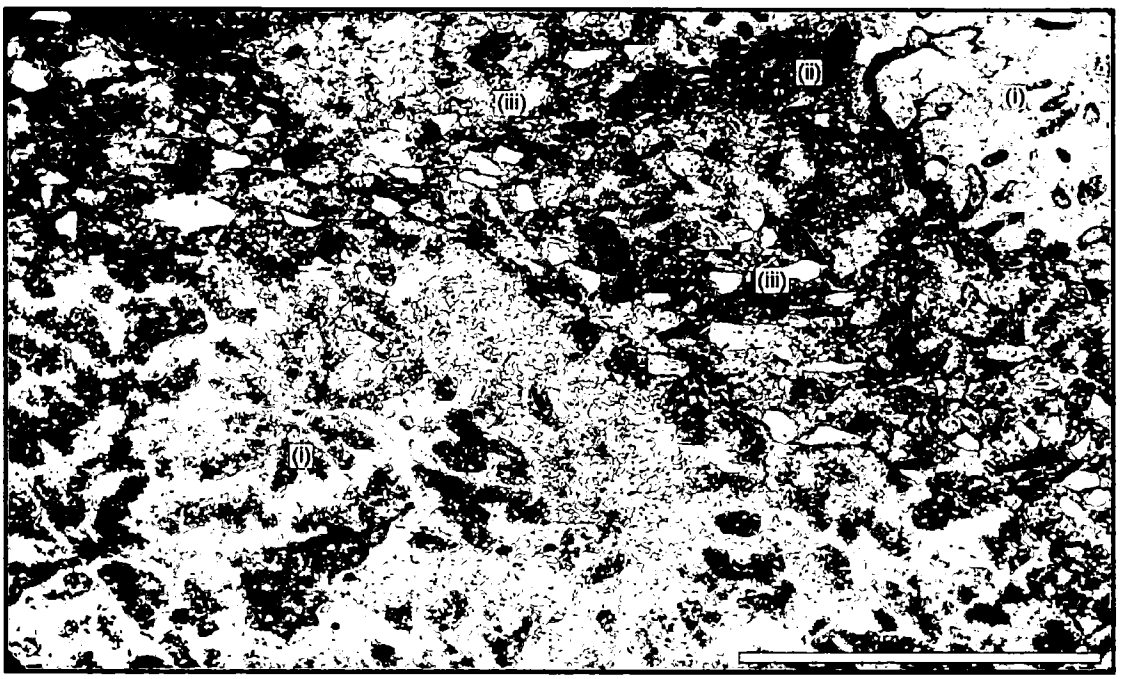


Figure 3.34a Photomicrograph (PPL) of the coralgal foraminifera siliciclastic float/rudstone facies. This facies is characterized by an abundance of corals (i) and coralline algae (ii) with abundant angular siliciclastic grains (iii) in the neomorphic microspar matrix. Note that corals are composed of a replacive drusy calcite spar cement (sample LCA 74, log CA-8). Scale bar=5 mm.

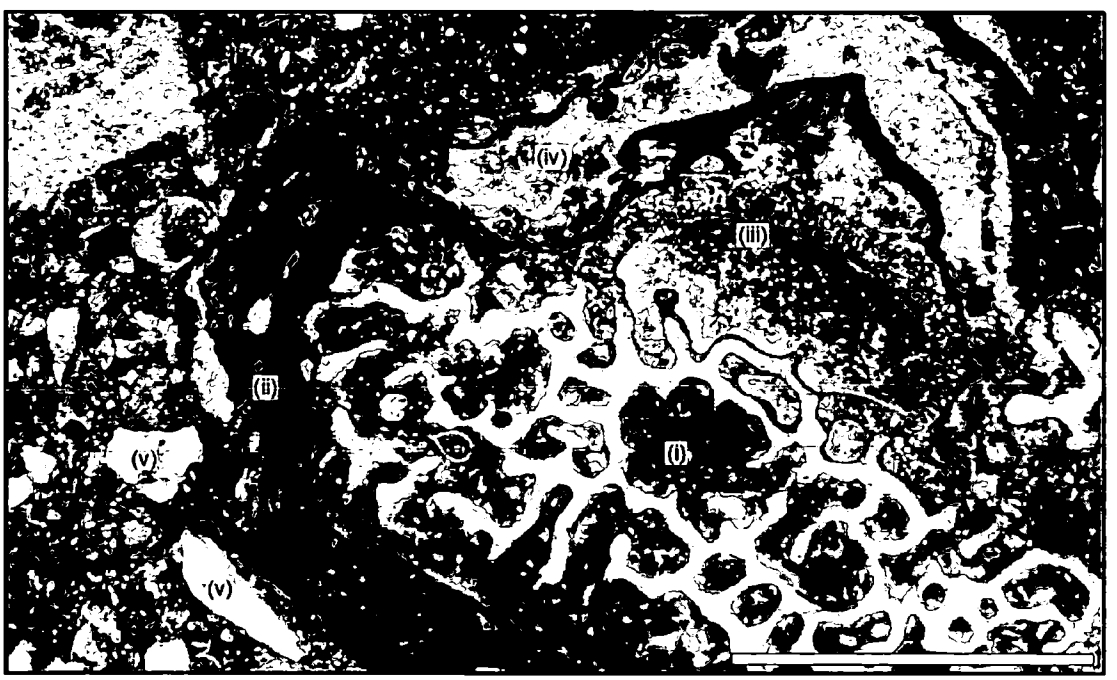


Figure 3.34b Photomicrograph (PPL) of the coralgal foraminifera siliciclastic float/rudstone facies. Rhodoliths are nucleated on coral fragments (i), and demonstrate thick, laminar coralline algae crusts (ii) with additional *Haddonia* (iii) and acervulinid crusts (iv) that have developed into short, stout branches. (sample LCA 43b, log CA-4a). Scale bar=2 mm.

stenohaline biota (Blondeau 1972, Chaproniere 1975, Murray 1991). Deposition within the photic zone is interpreted from the presence of symbiont-bearing larger foraminifera such as *Calcarina*, *Amphistegina* and *Nummulites*, and the presence of *in situ* coral colonies (Hottinger 1983, Hallock and Glenn 1985, Murray 1991, Hohenegger *et al.* 1999).

The significant percentage of non-carbonate material within samples of this facies indicates that this environment experienced a significant siliciclastic input. It is postulated from the presence of cryptic fauna (e.g. *Haddonia*) that corals were growing suprastratally, although there is no evidence to suggest the corals made up a framework. Colonies that stand proud of the sediment can influence the hydrodynamic conditions in their immediate vicinity, effectively trapping suspended sediment (Insalaco 1998). It is suggested that the significant percentage of fine-grained material is attributed to a combination of baffling by corals and/or low-energy conditions at times.

Although most biota are fragmented, minor abrasion indicates limited transport and reworking. The lenticular, robust morphologies of *Amphistegina* and *Nummulites* are characteristic of the shallower parts of their depth range in agitated waters (Leutenegger 1984, Hallock and Glenn 1985, Hallock 1988). Additionally, the presence of tight concentric rhodoliths (Bosence 1983a, 1983b, Minnery *et al.* 1985, Minnery 1990) and an abundance of *Gypsina, Calcarina* and miliolids may also be characteristic of shallow, agitated conditions (Chaproniere 1975, Reid and MacIntyre 1988, Hallock 1998, Murray 1991).

It is concluded that this environment experienced variable hydrodynamic conditions, with fragmentation of fauna during high-energy incursions. High-energy periods are associated with coarse siliciclastic input. Coarse siliciclastic input and unstable substrates may have inhibited widespread colonisation of this environment by corals.

In summary, the coralgal foraminifera siliciclastic float/rudstone facies is interpreted to represent patchy coral development within a periodically agitated, open marine environment experiencing a significant siliciclastic input. This facies was deposited in a similar environment to the coralgal foraminifera float/rudstone (see **Section 3.3.1.6**), but experienced a more significant siliciclastic input.

3.3.3 Bioclastic siliciclastic-dominated facies group

Facies included in this grouping are predominantly siliciclastic in composition, although they may contain a significant carbonate component as a combination of bioclasts, cement and detrital carbonate grains. The dominant texture is sandstone, and grains are well-sorted, sub-angular and sub-spherical. Biota are dominated by benthic foraminifera including *Gypsina,* Nummulites and *Amphistegina*, with molluscs, brachiopods, echinoids and serpulids (**Table 3.1**).

3.3.3.1 *Gypsina* calcarenite

Lithologies: *Gypsina* calcarenite
Gypsina siliciclastic pack/wackestone

Occurrence and Bed Characteristics

The *Gypsina* calcarenite facies is encountered in the upper part of the Calders succession between carbonate intervals 5 and 6 (**Figure 3.9**). This facies occurs as planar cross-stratified beds that may be traced laterally for over 1 km. Set height is up to 5 m, and foresets dip consistently towards the north. Measured bed thickness ranges from 1 m to 2.5 m, and beds can be traced laterally for at least 1 km. Upper and lower bedding contacts are sharp and non-erosional. Bioturbation includes vertical and horizontal *Skolithos* and *Ophiomorhpa*-type burrows.

Lithological description

The weathered surface of the foraminifera litharenite facies ranges from buff to dark grey in colour. The total non-carbonate content of this facies ranges from 35.4 to 69 wt. %. Non-carbonate grains range from clay to coarse sand grade (**Figure 3.35a**). Maximum siliciclastic grainsize is 0.5 mm. Coarse grains are angular and moderately well-sorted (**Figures 3.35a** and **b**). Siliciclastic grains are predominantly mono- and polycrystalline quartz (up to 20.5 %), lithic grains (8.5 to 15 %) with orthoclase feldspar (up to 4 %) (**Figure 3.35b**). Lithic grains are siltstones and quartz-mica schist. The micas have commonly weathered to a yellow to rusty-brown coloured clay observed in thin section and orthoclase feldspar grains have a brown altered appearance (**Figure 3.35b**).

Carbonate material comprises up to 64 % of this facies as a combination of bioclasts, detrital calcite grains, cement and micrite matrix. Micrite comprises up to

28.5 % of this facies. This facies has a moderately diverse foraminifera assemblage dominated by *Gypsina* (up to 21 %) with *Rotorbinella* (0 to 7.5 %), *Amphistegina* (0 to 1.5 %), *Fabiania* (up to 3 %), *Calcarina* (< 0.5 %), miliolids (0 to 1.5 %) and *Nummulites* (0 to 3.5 %). *Gypsina* tests are rounded in thin section (**Figure 3.35a**). Robust lenticular forms of *Amphistegina* and *Nummulites* tests are up to 0.8 mm and are fragmented and abraded. Miliolids are small (less than 0.4 mm) and are abraded. *Fabiania* tests up to 2 mm occur intact and as abraded fragments (**Figure 3.35b**). *Rotorbinella* tests are always fragmented (**Figure 3.35a**).

Other biota present in this facies are coralline algae (0 to 11.5 %), peloids (1 to 7.5 %), echinoids (0 to 7.5 %), bryozoa (0 to 3.5 %), molluscs (0 to 2 %) and micritic serpulid tubes (0 to 2.5 %) (**Figures 3.35c** and **d**). Coralline algae are present as abraded laminar fragments with poorly preserved internal structure. Peloids are very well sorted and are interpreted as heavily abraded coralline algae fragments.

Diagenesis

Gypsina tests with a rounded morphology occasionally have thin micrite envelopes (**Figure 3.35b**). Intragranular microspar cement is observed within foraminifera tests and echinoid spines demonstrate well-developed syntaxial calcite cements. Originally aragonitic molluscan fragments have been leached with the resultant biomouldic pores completely occluded by clear, drusy calcite cement. Patchy neomorphic microspar (5 to 10 %) is present in intergranular areas.

Interpretation: depositional environment

The *Gypsina* calcarenite facies contains a marine faunal assemblage. Deposition is inferred to have taken place within the photic zone from the presence of symbiont-bearing benthic foraminifera such as *Amphistegina*, *Nummulites* and *Calcarina* although some transport and reworking of grains is deduced (Leutenegger 1984, Hallock and Glenn 1985, 1986; Murray 1991, Hohenegger *et al.* 1999).

A moderate siliciclastic input, and consequent unstable substrates, is inferred from the abundance of siliciclastic material in this facies. The sand grade component would have been transported as bedload with finer-grained material deposited from suspension. Limited abrasion due to transport or reworking prior to deposition and lithification is interpreted from the angular nature of the siliciclastic grains. Moderate-

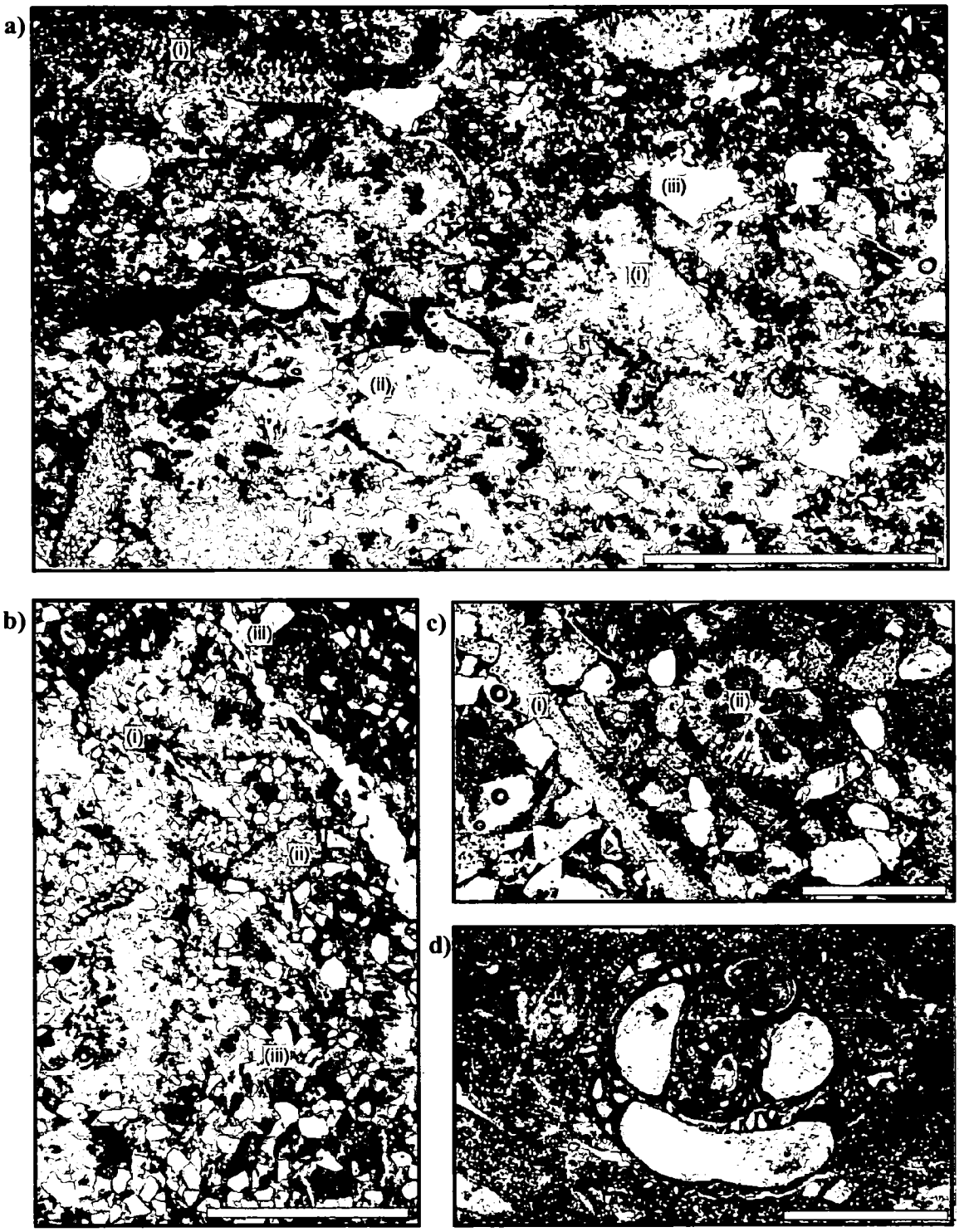


Figure 3.35 Photomicrographs (all PPL) of the *Gypsina* calcarenite facies. **a)** *Gypsina* tests (i) associated with fragmented *Rotorbinella* (ii). Siliciclastic grains (iii) up to coarse sand-grade are well-sorted (sample LCA 49, log CA-4 bed 23) Scale bar=2mm. **b)** Intact *Fabiania* (i) test. *Gypsina* tests (ii) show incipient micrite envelopes. Orthoclase feldspar (iii) is identified through its brown colour in thin section (sample LCA 47, log CA-4a bed 21). **c)** Echinoids (i) and bryozoa are present in this facies (sample LCA 99a, log CA-10 bed 6). Scale bar=1 mm. **d)** Intact serpulid tube, interpreted to have encrusted sea grass or macroalgae (sample LCA 49, log CA-4a bed 21). Scale bar=2mm.

energy, agitated conditions are consistent with the presence of coralline algae peloids (Wolf 1965, Friedmann *et al.* 1973).

The robust lenticular morphology of *Amphistegina* is indicative of the shallow to intermediate part of its depth range in higher energy/agitated conditions (Hottinger 1983, Hallock and Glenn 1985, 1986). The presence of sea grass is inferred from the abundance of *Gypsina, Amphistegina, Calcarina* and miliolids (Chaproniere 1975, Ghose 1977, Parsons-Hubbard *et al.* 1998, Hohenegger *et al.* 1999, Walker 2001). It is postulated that the baffling effect of sea grass may have trapped suspended sediment. The foraminifera assemblage characteristic of this facies contains foraminifera that often have an epiphytic lifestyle. It is suggested that foraminifera that were able to adapt an epiphytic lifestyle had an advantage in this environment as siliciclastic input and unstable substrates would have inhibited colonisation by most calcareous benthic organisms. Strictly benthic foraminifera (i.e. *Nummulites*) are relatively rare.

In summary, it is interpreted that the *Gypsina* calcarenite facies represents deposition in a siliciclastic-dominated, agitated, shallow marine environment with unstable substrates. The *Gypsina* calcarenite facies is comparable to the back-reef platform sediments of the Miocene Ziqlag Formation, Israel (Buchbinder 1977) and the Quartzitic Grainstone facies of Sayer (1995). Comparable foraminifera associations are identified in sandstones of Eocene outcrops in the northern Vic Basin (Alvarez *et al.*, 1995; Franquès-Faixa 1996). These sediments have been interpreted to represent a back-reef and/or lagoonal depositional setting.

3.3.3.2 *Nummulites* arenite

Lithologies: *Nummulites* arenite

Occurrence and bed characteristics

The *Nummulites* arenite facies is only encountered underlying the first carbonate interval (**Figure 3.9**). Measured bed thickness ranges from 1.15 m to 1.65 m, and can be traced laterally over 500 m. Upper and lower bedding contacts are gradational. Bioturbation is evident in the form of slight blue mottling of fresh surfaces of this facies.

Lithological description

The surface of the *Nummulites* arenite facies weathers to a reddish-brown colour. Fresh surfaces are a light brown to buff colour. This facies is composed of well-sorted, angular to sub-rounded monocrystalline quartz (25.5 to 40 %), lithic (2.5 to 5 %) and orthoclase feldspar (up to 1 %) grains situated within a neomorphic microspar matrix (**Figure 3.36**). Maximum siliciclastic grainsize is 1.75 mm. Mean siliciclastic grainsize is 0.5 mm. Lithic grains are siltstone and quartz-mica schist. The micas have commonly weathered to rusty-brown coloured clay observed in thin section. Orthoclase feldspar grains have a brown, altered appearance in thin section.

The *Nummulites* arenite facies contains a low diversity foraminifera assemblage dominated by intermediate lenticular forms of *Nummulites* (2.5 to 4 %) with *Gypsina* (1.5 %) and miliolids (0.5 %) (**Figure 3.36**). *Nummulites* test size ranges from 0.5 mm to 4 mm. Tests are typically fragmented and fragments are abraded. Laminar *Gypsina* tests up to 3 mm in length are fragmented and abraded (**Figure 3.36**). Miliolids are abraded with slightly reddened tests. Subordinate bioclasts within the *Nummulites* arenite facies are fragmented echinoid spines (< 0.5 %), bryozoa (< 0.5 %) and brachiopod spines (< 0.5 %). Fragmented bioclasts are angular and poorly sorted.

Diagenesis

The originally micritic matrix of the *Nummulites* arenite facies has recrystallised into microspar (**Figure 3.28**). Intragranular calcite cement is observed within foraminifera chambers and echinoid spines show poorly developed syntaxial calcite overgrowth cements.

Interpretation: depositional environment

The *Nummulites* arenite facies contains a marine faunal assemblage. Deposition in a shallow marine environment within the photic zone is inferred from the modest abundance of *Nummulites* (Ghose 1977, Hallock and Glenn 1986, Hallock 1988). The intermediate, lenticular morphology of *Nummulites* is indicative of the shallow to intermediate part of its habitat range in moderate energy conditions (Hallock and Glenn 1985, 1986). It is suggested from the fragmented and slightly abraded nature of foraminifera that they have undergone *in situ* reworking. Water agitation was probably sufficient to remove smaller bioclasts.

Rapid deposition prior to reworking and abrasion is inferred from the angular nature of grains. It is postulated that high siliciclastic sedimentation rates and/or unstable substrate inhibited colonisation of the sediment by calcareous benthic organisms. This is supported by the scarcity of fauna in this facies. Variable depositional energies are inferred from the locally present micrite matrix (that has transformed into neomorphic spar). It is suggested that *Nummulites* colonized this environment during relative quiet periods.

In summary, the *Nummulites* arenite facies formed within a shallow marine, siliciclastic shelf setting characterised by unstable substrates. Similar poorly fossiliferous facies are described from the Nummulitique (Sayer 1995). These sediments are interpreted as shallow marine offshore sand shoals and bars.

3.3.3.3 Bioclastic siltstone

Lithologies: Bioclastic mudstone
Bioclastic muddy litharenitic siltstone

Occurrence and bed characteristics

The bioclastic siltstone facies is only encountered within carbonate interval 5 (**Figure 3.9**). Measured bed thickness is 0.4 m, although the lateral extent cannot be determined due to poor exposure. Upper and lower bedding contacts are sharp and non-erosional.

Lithological description

Exposed surfaces of the bioclastic siltstone facies weather to a buff colour. Fresh surfaces are a grey-brown colour with blue-grey patches. This facies contains clay to silt-grade siliciclastic grains floating within a muddy matrix (**Figure 3.37**). Siliciclastic grains are mono- and polycrystalline quartz (5 %) with lithics (0.5 %) and orthoclase feldspar (0.5 %). Lithic grains are micaceous siltstones. The micas have commonly weathered to rusty brown clay. Grains are angular with a low sphericity, and are very well sorted. The matrix of the bioclastic siltstone facies is a homogenous dark brown colour in thin section. The matrix is composed of micrite (up to 47.5 %) and non-carbonate clay (up to 40 %).

Poorly sorted bioclasts constitute a relatively minor percentage of this facies, and include intermediate *Nummulites* (up to 2 %) and robust, lenticular *Amphistegina*

(0.5 %). *Nummulites* tests up to 4 mm in diameter are preserved intact and slightly abraded (**Figure 3.37**). Other bioclasts found in this facies are *Gypsina* (0.5 %), corals (0.5 %), echinoid spines (1 %), bivalves (2 %) and coralline algae (0.5 %). Delicate bivalves are preserved articulated. Coralline algae fragments are highly abraded, and little internal structure is preserved. Bioclasts may be concentrated within small *Skolithos*-type vertical burrows (**Figure 3.37**).

Diagenesis

Intragranular microspar cements are observed within foraminifera tests, and originally aragonitic bioclasts have been leached with the mouldic pores completely cemented with calcite spar. Patchy recrystallisation of the originally micritic matrix into microspar (2 %) is also evident.

Interpretation: depositional environment

The bioclastic siltstone facies contains a marine faunal assemblage. Normal marine conditions are inferred from the stenohaline biota (Chaproniere 1975, Hallock and Glenn 1986, Murray 1991, Geel 2000). Low-energy conditions are inferred from the fine-grained nature of this facies. It is interpreted that clay to silt-grade siliciclastic material was deposited from suspension. Limited transport and abrasion prior to deposition and lithification is inferred from the angular nature of siliciclastic grains.

Deposition within the photic zone is inferred from the presence of *Nummulites* and *Amphistegina* (Leutenegger 1984, Hallock and Glenn 1985, Murray 1991). The intermediate morphology of *Nummulites* and *Amphistegina* is indicative of the intermediate part of their habitat range within moderate energy conditions (Hottinger 1983, Hallock and Glenn 1985, 1986). It is postulated that abraded bioclasts and foraminifera tests have been washed-in from a relatively higher energy environment.

In summary, the bioclastic siltstone facies accumulated in an open marine, low energy setting with a moderate siliciclastic and bioclastic input from high-energy parts of the shelf. This facies has affinities with the Foraminiferal Mudstone facies of Sayer (1995). These sediments were deposited within a very low-energy, offshore environment unaffected by shallow marine currents.

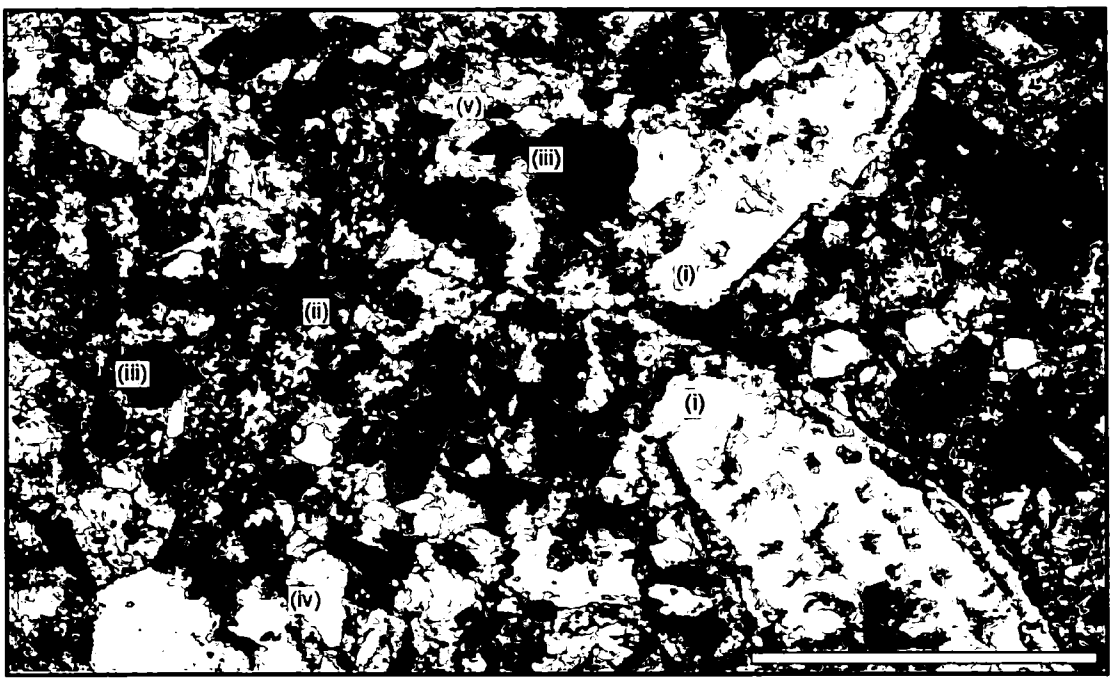


Figure 3.36 Photomicrograph of the *Nummulites* arenite facies. This facies contains large, lenticular *Nummulites* (i) and laminar *Gypsina* (ii) tests. Highly abraded coralline algae fragments (iii) are also present (sample LCA 17, log CA-2 bed 1). Scale bar=1 mm

Figure 3.37 Photomicrograph of the bioclastic siltstone facies. Bioclasts including *Nummulites* (i), coralline algae (ii) and echinoid spines (iii) occur aligned in burrows (sample LCA 46, log CA-4a). Scale bar=2 mm.

3.3.4 Siliciclastic-dominated facies group

Facies included in this grouping are predominantly siliciclastic in composition, although they may contain a moderate carbonate component as a combination of cement and detrital carbonate grains. The dominant texture is sandstone, and grains are well to poorly sorted, sub-angular and sub-spherical. This facies grouping differs from the bioclastic siliciclastic-dominated facies group in that bioclasts are very rare, although bioturbation is typically evident (**Table 3.1**).

3.3.4.1 Cross-stratified calcarenite

Lithologies: Calcarenite
Muddy calcarenite
Calcarenitic siltstone
Interbedded calcareous litharenitic silt and sandstone

Occurrence and bed characteristics

This facies occurs throughout the succession (**Figure 3.9**). Measured bed thickness ranges from less than 10 cm to 3 m. Beds of the cross-stratified calcarenite facies have a planar to low-angle sigmoidal morphology, and are characterised by metre-scale cross-stratification (**Figures 3.38a** and **b**). Foresets and topsets are preserved, with set height ranging from 1 to 5 m (**Figures 3.38a** and **c**). Cross-sets show clear progradational morphology, with foresets dipping 15° to 35° towards the north. Upper and lower bed contacts are typically sharp and non-erosional. The upper contact with the lenticular polymict conglomerate facies is always erosional (**Figure 3.39a**).

Swaley cross stratification is observed in some localities (**Figure 3.39a**). Symmetrical ripple lamination is superimposed on swaly cross-beds. Small channel structures are observed in the cross-stratified calcarenite facies (**Figure 3.38c**). Channels have an erosional base, and channel in-fills include lateral accretion surfaces. Siltstone and shale interbeds, up to 10 cm in thickness, comprise up to 10 % of this facies (**Figure 3.38c**). Vertical *Skolithos* and *Ophiomorpha*-type burrows are present locally. Burrows do not cross upper bedding contacts.

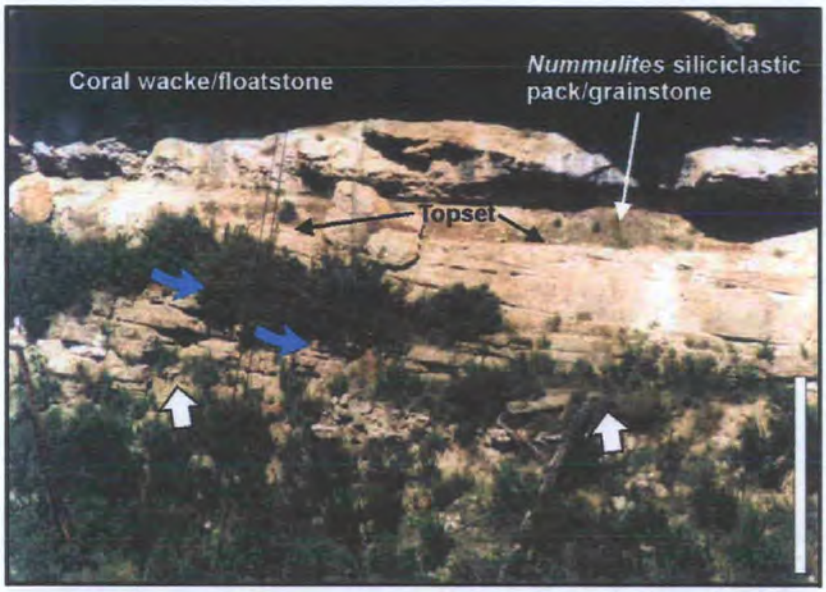


Figure 3.38a Cliff exposure of metre scale cross-stratification within the cross-stratified calcarenite facies. Set height is approximately 5 m. The progradation direction (blue arrows) is oblique into the cliff face (broadly northward). The *Nummulites* siliciclastic pack/ grainstone facies was deposited on the distinct topset surface (indicated). Photograph taken from GR 16902625, facing NNW. Scale bar=5 m approx.

Figure 3.38b Low-angle sigmoidal bedding within the cross-stratified calcarenite facies (log CA-7 beds 51 to 52). Facing NE. Scale bar=2 m approx.

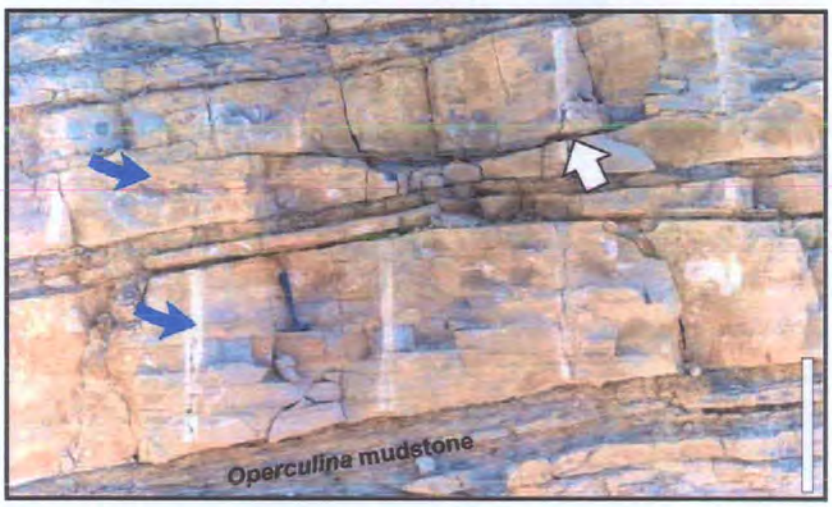


Figure 3.38c Interbedded sandstones and grey shales. Note the dm-scale cross-stratification (blue arrows indicating flow put of the page) and the small channel structure (white arrow). The channel infill shows lateral accretion surfaces. Photograph taken from GR 16752675, facing SE. Scale

Apparent palaeoflow/progradation direction

Lithological description

Fresh surfaces of this facies are a pale-brown to buff colour. Weathered surfaces are a dark brown colour. Grains are angular, well sorted and subspherical (**Figures 3.39b** and **c**). Grainsize ranges from clay to coarse sand. Siliciclastic grains identified are monocrystalline strained and unstrained quartz (20 to 46 %), lithics (5 to 45.5 %), feldspar 0.5 to 2.5 %) and muscovite mica (up to 2 %). Lithic grains are siltstones, quartz-mica schist and compacted clay grains (**Figure 3.39c**). Micas commonly have a 'rotted' appearance, and are altered to a yellow to rusty-brown coloured clay. Detrital carbonate (calcite and dolomite) comprises 2 to 30 % of grains. Calcite grains and dolomite rhombs are interpreted to be reworked cements. Dolomite is only abundant in sample LCA 229 (log CA-15 bed 10) (**Figure 3.39c**). The matrix of this facies is composed of a combination of micrite (6 to 20 %) and non-carbonate clays (1.5 to 30 %).

The cross-stratified calcarenite facies is largely unfossiliferous. Large irregular echinoids up to 8 cm are preserved intact (log CA-4a, bed17). In addition, *Cerithium* gastropods up to 45 cm in length are abundant in one horizon (log CA-5, bed 7). Subordinate biota present are fragmented echinoid spines (0 to 1.5 %), *Gypsina* (0 to 2 %), miliolids (0 to 1 %) and coralline algae peloids (0 to 5 %). Bioclasts are highly abraded.

Diagenesis

Intergranular areas have been cemented with equant calcite spar (up to 40 %). Originally aragonitic biota have been leached, although biomouldic macropores have not been occluded. Echinoid spines may show syntaxial calcite overgrowth cements.

Interpretation: depositional environment

Deposition of this facies in a marine environment is concluded from the marine biota and presence of symmetrical ripple lamination. Metre-scale cross-stratified beds are interpreted as subaqueous dunes (Reading 1996). Deposition at depths above fair weather wave base with a strong current influence is interpreted from the presence of dunes. It is postulated that mudstone and shale interbeds were deposited either during relatively low-energy periods or in inter-dune areas. Periodic deposition between storm and fair weather wave base in the offshore transition zone is concluded from the presence of swaley cross-stratification (Reading 1996).

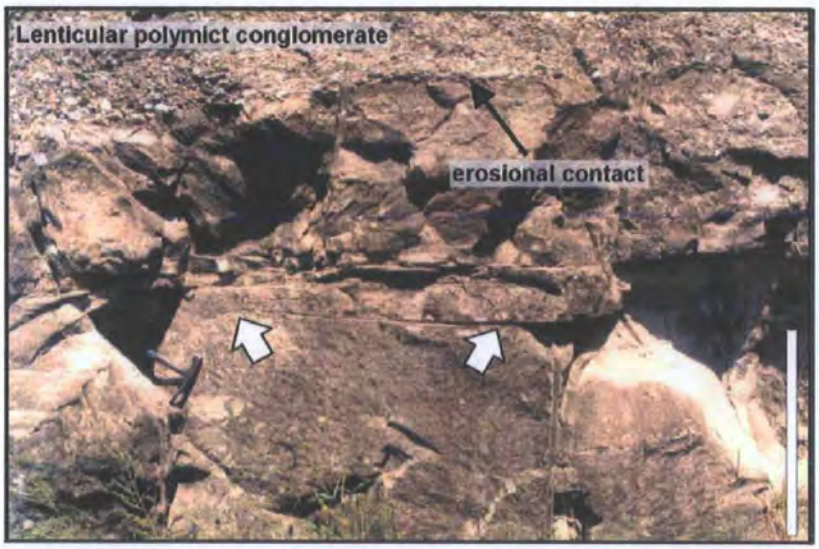


Figure 3.39a Possible swaley cross-stratification of the cross-stratified calcarenite facies (arrowed). The upper contact of the sandstone with the lenticular polymict conglomerate is erosional (indicated). (log CA-7 beds 11 to 13). Scale bar=1 m approx.

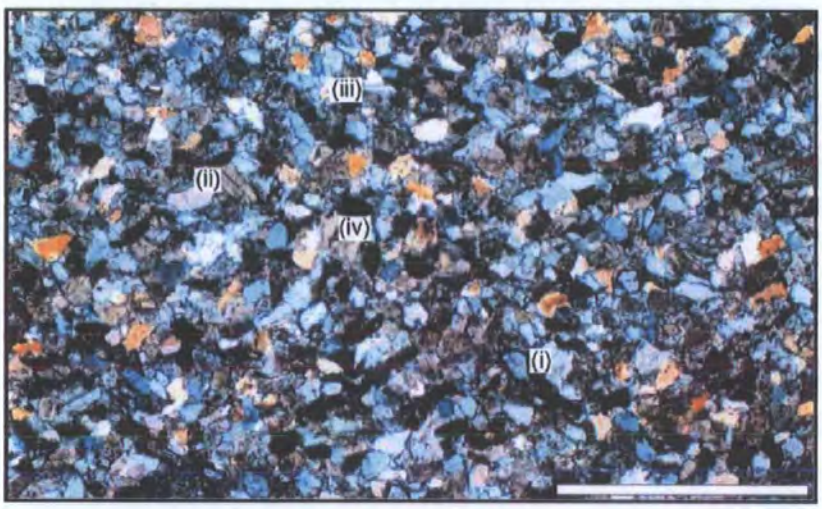


Figure 3.39b Photomicrograph (XPL) of the cross-stratified calcarenite facies. This facies is dominated by quartz (i) with calcite (ii) and feldspar (iii). Highly abraded coralline algae fragments (iv) are present in some samples (sample LCA 21, log CA-3 bed 6). Scale bar=2 mm.

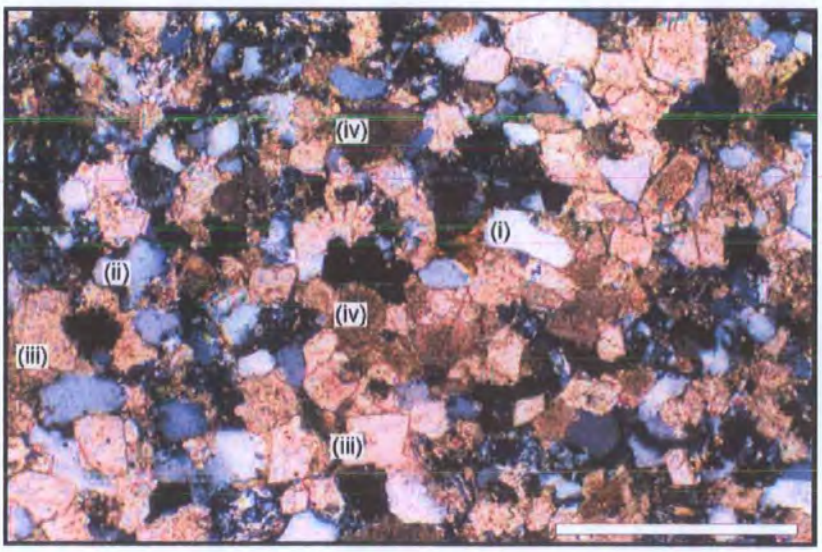


Figure 3.39c Photomicrograph (XPL) of a dolomite-rich calcarenite. This sample contains mono-crystalline unstrained and strained quartz (ii), with detrital dolomite (iii) and compacted mud clasts (iv). Dolomite rhombs are inferred to be detrital in origin from the presence of circum-granular clay coatings and the presence of fragmented and abraded rhombs (sample LCA 229, log CA-15 bed 10). Scale bar= 0.5 mm

Limited abrasion of siliciclastics prior to deposition and lithification is inferred from the angular nature of grains, although the grains are very well-sorted. Rapid deposition is inferred from the sharp nature of contacts and the largely structureless nature of beds. Rapid deposition of sediment is also suggested from the truncation of burrows at bedding contacts. Only mobile calcareous biota (gastropods and echinoids) are preserved *in situ* in this facies. Foraminifera and coralline algae are highly abraded and thus have been reworked from another facies. It is suggested that a consequence of rapid sedimentation there was an unstable substrate that inhibited colonisation by sessile calcareous benthic organisms. Silt and shale interbeds are interpreted to have been deposited during relatively quiet periods. A progradational sedimentary regime is concluded from the cross-stratifcation of beds of this facies. Progradation direction is consistently towards the north.

In summary, the cross-stratified calcarenite facies represents siliciclastic-dominated deposition within a marine setting at depths ranging from above fair weather base down to storm wave base. High sedimentation rates and unstable substrates inhibited the development of a sessile calcareous benthic community. Matrix-rich examples of this facies are comparable with the Quartzitic Wackestone facies of Sayer (1995). Well-cemented examples, characterised by negligible fine-grained matrix, are comparable with the Quartzitic Grainstone facies of Sayer (1995). Sayer (1995) attributes the presence/absence of matrix to variable water energies. The Quartzitic Wackestone and Quartzitic Grainstone facies have been interpreted as offshore shoal and bar-type environments reworked by continual wave and current action.

3.3.4.2 Lenticular polymict conglomerate

Lithologies: Polymict conglomerate

Occurrence and bed characteristics

The lenticular polymict conglomerate facies weathers recessively and is only exposed in cut sections along the B124 road (**Figure 3.1b**). This facies occcurs as lenticular sedimentary bodies that are traced 15 to 20 m along strike (**Figure 3.40**) and up to 100 m down-dip. Measured bed thickness is up 1.4 m. The lower bed contact with calcarenitic sands is erosional. Scours up to 15 cm infilled with massive

conglomerate are common (**Figure 3.40**). Steep cross-stratification is observed (10° to 45°), although the lenticular polymict conglomerate is largely lacking internal structure (**Figures 3.41a** and **b**).

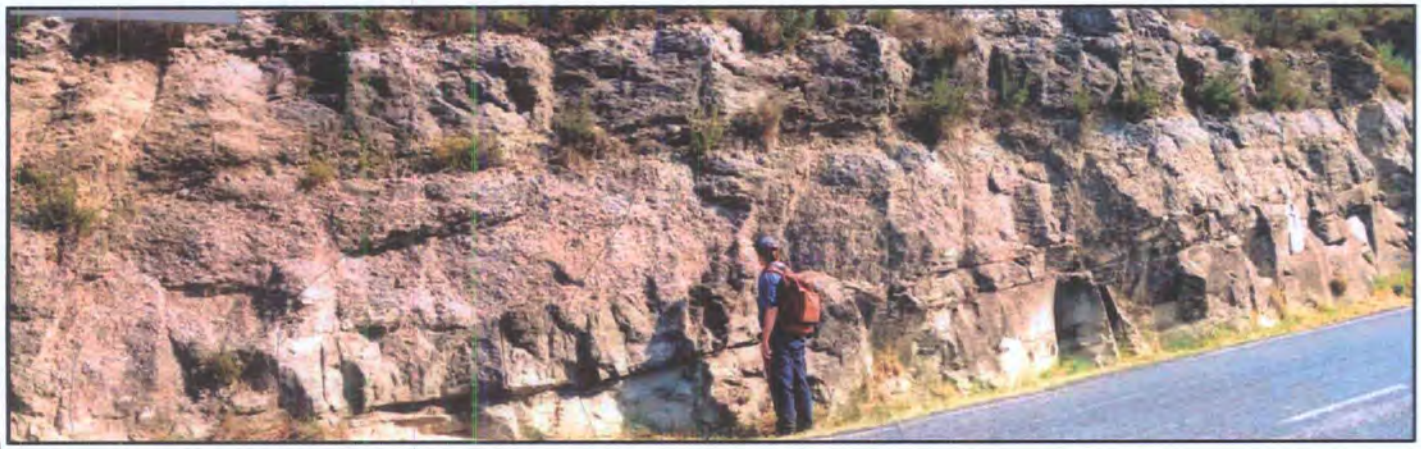
Lithological description

The lenticular polymict conglomerate facies weathers to a grey to dark red-brown colour. This facies is a matrix-supported conglomerate. Clasts within this facies range from clay-grade up to 7 cm in diameter. Clasts are sub-angular to very well rounded, poorly sorted with low to moderate sphericity. Clasts consist of quartz (up to 45 %), lithics (up to 50 %) with minor feldspar (less than 5 %). Lithic grains are quartzite, chert, siltstone, sandstone and schists. Matrix consists of muddy, coarse-sand grade calcarenite. Matrix comprises up to 40 % of this facies. Bioclasts are largely absent within the lenticular conglomerate facies. Reworked spherical *Nummulites* tests up to 0.4 cm in diameter occur along the upper limits of this facies. *Nummulites* are largely intact but highly abraded.

Interpretation: depositional environment

The lenticular polymict conglomerate facies contains a very low diversity marine fauna. The small, robust forms of *Nummulites* are abundant in the shallow, high-energy part of their habitat range in the upper portions of the photic zone (Hallock and Glenn 1985, Murray 1991). Deposition in an environment close to the sediment source is inferred from the coarse-grained, poorly sorted nature of sediments. High-energy deposition is inferred from the erosive nature of the base of this facies and the coarse clast size. The steep cross stratification surfaces are orientated perpendicular to the main progradation direction of the underlying cross-stratified calcarenite facies. These surfaces are interpreted as lateral accretion surfaces, thus the lenticular conglomerate facies is interpreted as a high-energy channel-fill deposit. Deposition in a marine setting is inferred from the association of this facies with the cross-stratified calcarenite and *Nummulites* siliciclastic pack/grainstone facies.

a)



b)

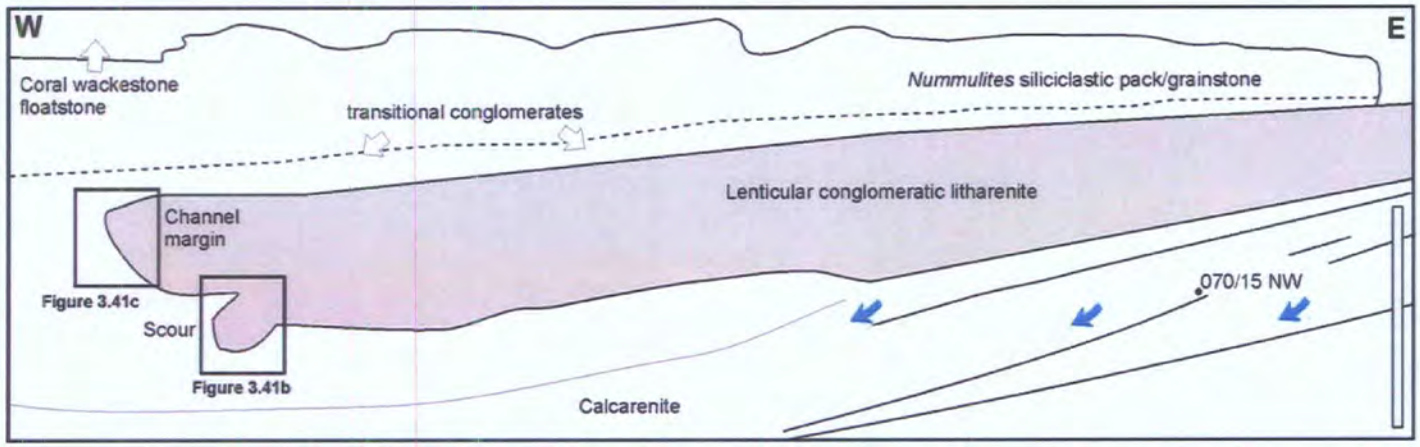


Figure 3.40 Roadcut exposure (a) and interpretation (b) of a lenticular polymict conglomerate unit, interpreted as a conglomerate-filled channel. Channel structures can be traced laterally for > 20 m. The lower contact with the cross-stratified calcarenite fcaies is erosional. Erosional scours up to 35 cm deep, are infilled with massive conglomerates. Photograph taken from GR 16752609, facing ENE. Scale bar=2 m approx.

Figure 3.41a Cross-stratification within the lenticular polymict conglomerate facies. Foresets (arrowed) are interpreted as lateral accretion surfaces (log CA-7, bed 14, facing N). Scale bar= 50 cm.

Figure 3.41b Massive bedded conglomerates (log CA-7, bed 14). Scale bar= 50 cm.

Figure 3.41c Margin of a lenticular conglomerate unit (arrowed). The contact between the conglomerate and underlying calcarenite is erosional. (log CA-7, bed 14). Scale bar= 50 cm.

3.4 Calders facies associations and depositional models

Detailed logging and log correlation has made it possible to reconstruct the temporal and spatial facies variations in the Calders area, and to examine the relationship between carbonate production and siliciclastic sedimentation. Sixteen sections in total were logged and sampled in the Calders area (**Figure 3.1** and **Appendix 5**). Six main carbonate intervals are identified in the Calders succession (**Figure 3.9**). Each carbonate interval, coupled with a siliciclastic interval, forms a carbonate-siliciclastic cycle. The only exception is cycle 5 that contains an additional carbonate interval (interval 5a) in the east of the study area (**Figure 3.44**). There are 6 full carbonate-siliciclastic cycles. Cycles 1 to 4 are represented by logs in the southern parts of the Calders study area, with cycles 4 to 6 represented by logs in the northern area (**Figure 3.42**). Facies correlations of the southern and northern parts of the Calders study area are presented in **Figures 3.43** and **Figure 3.44** respectively.

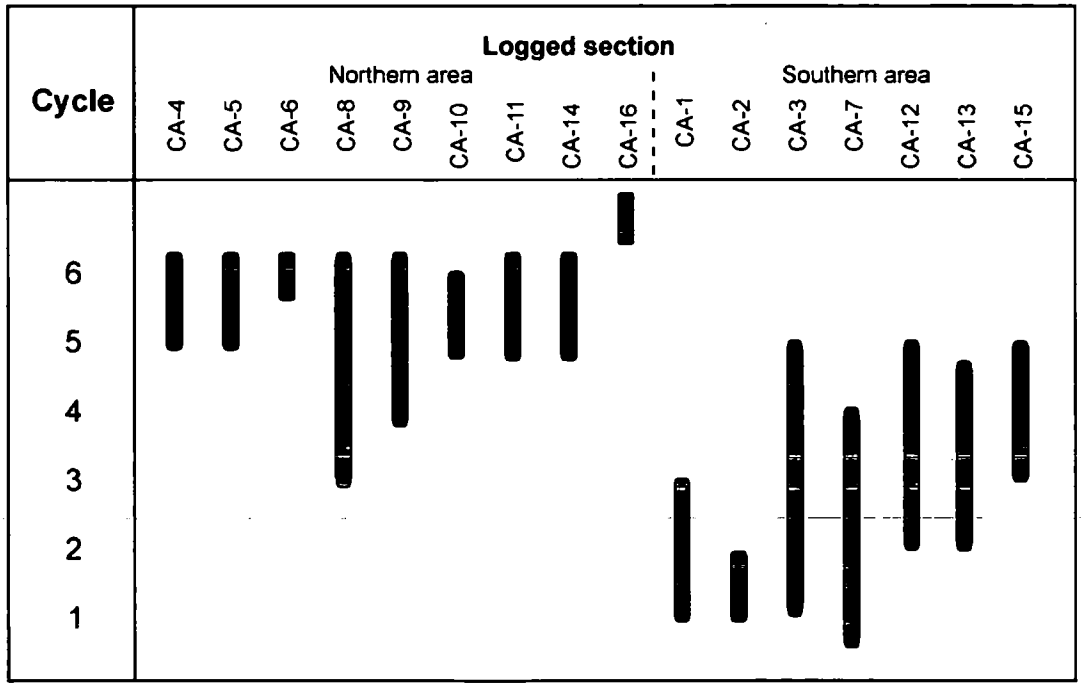


Figure 3.42 Coverage of carbonate-siliciclastic cycles by logged sections. Logged sections in the northern roughly cover cycles 4 to 6. Logged sections in the southern area cover cycles 1 to 4. Log CA-16 represents siliciclastic sediments deposited stratigraphically above cycle 6 and below the Sant Amanc succession (discussed in **Section 3.5**)

Figure 3.43 a) Facies correlation.

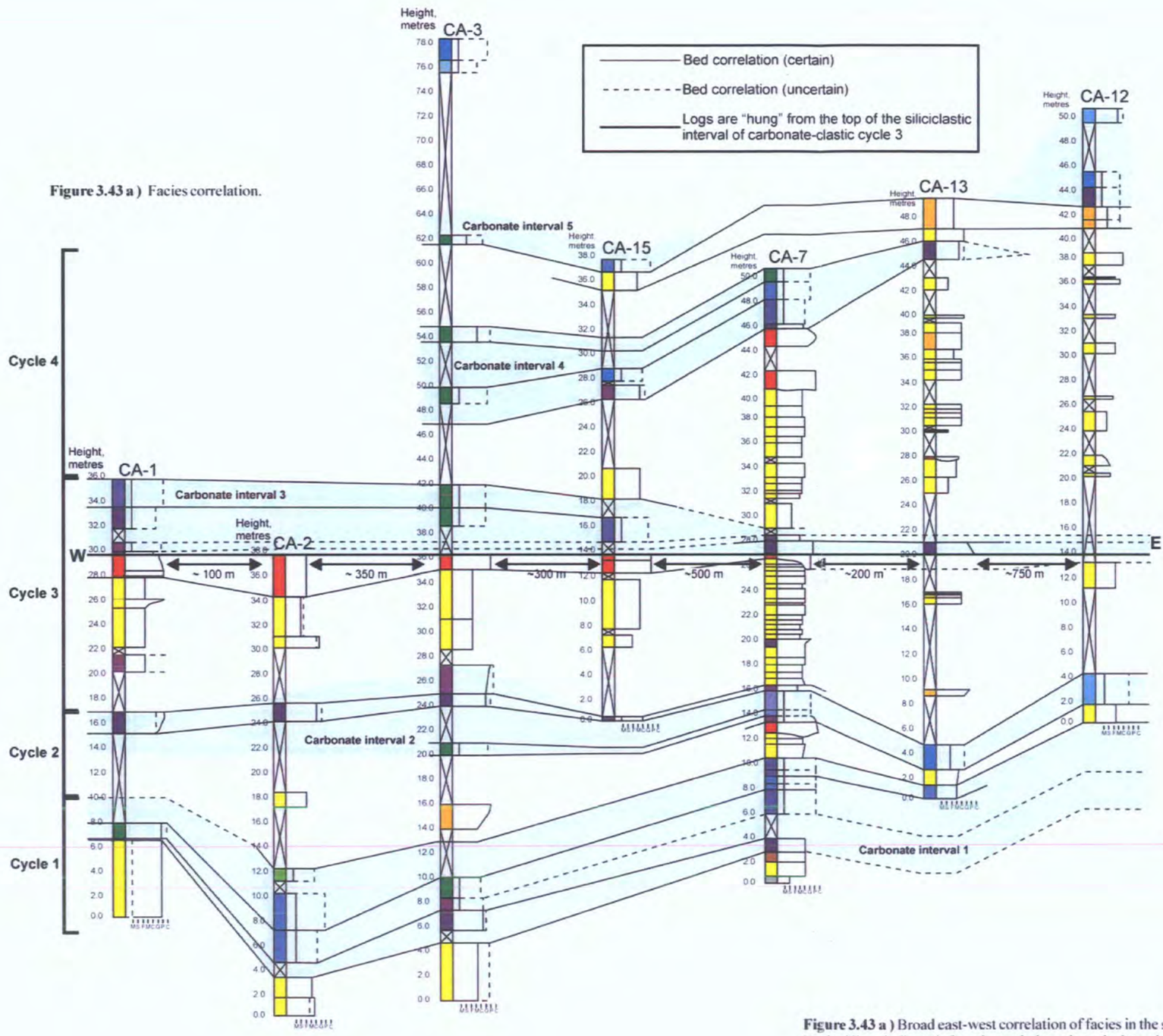
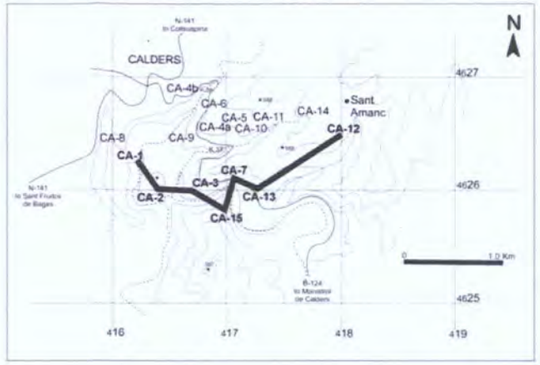


Figure 3.43 b) Sketch location map



Facies key

- Coral mixstone (**Section 3.3.1.1**)
- Coral domestone (**Section 3.3.1.2**)
- Coral bioclastic pack/rudstone (**Section 3.3.1.3**)
- Coral wacke/floatstone(**Section 3.3.1.4**)
- Coralgal foraminifera rudstone (**Section 3.3.1.5**)
- Coralgal foraminifera float/rudstone (**Section 3.3.1.6**)
- Foralgal grainstone (**Section 3.3.1.7**)
- *Gypsina* grainstone (**Section 3.3.1.8**)
- Miliolid coralline algae pack/grainstone (**Section 3.3.1.9**)
- *Nummulites* siliciclastic pack/grainstone (**Section 3.3.2.1**)
- *Nummulites Discocyclina* siliciclastic packstone (**Section 3.3.2.2**)
- *Operculina* mudstone (**Section 3.3.2.3**)
- *Nummulites* coralline algae siliciclastic pack/grainstone (**Section 3.3.2.4**)
- *Gypsina* calcarenite (**Section 3.3.3.1**)
- *Nummulites* arenite (**Section 3.3.3.2**)
- Bioclastic siltstone (**Section 3.3.3.3**)
- Cross stratified calcarenite (**Section 3.3.4.1**)
- Lenticular polymict conglomerate (**Section 3.3.4.2**)

Figure 3.43 a) Broad east-west correlation of facies in the southern part of the Calders study area. All slid tie-lines have been traced in the field. Dashed lines indicate inferred correlation where the lateral exposure of beds is poor. Carbonate intervals are shaded in pale blue for clarity. Each carbonate-siliciclastic cycle is indicated. Logs are not arranged to scale horizontally. A sketch map showing locations of logged sections is present in **b)**.

Figure 3.44 a) Facies correlation.

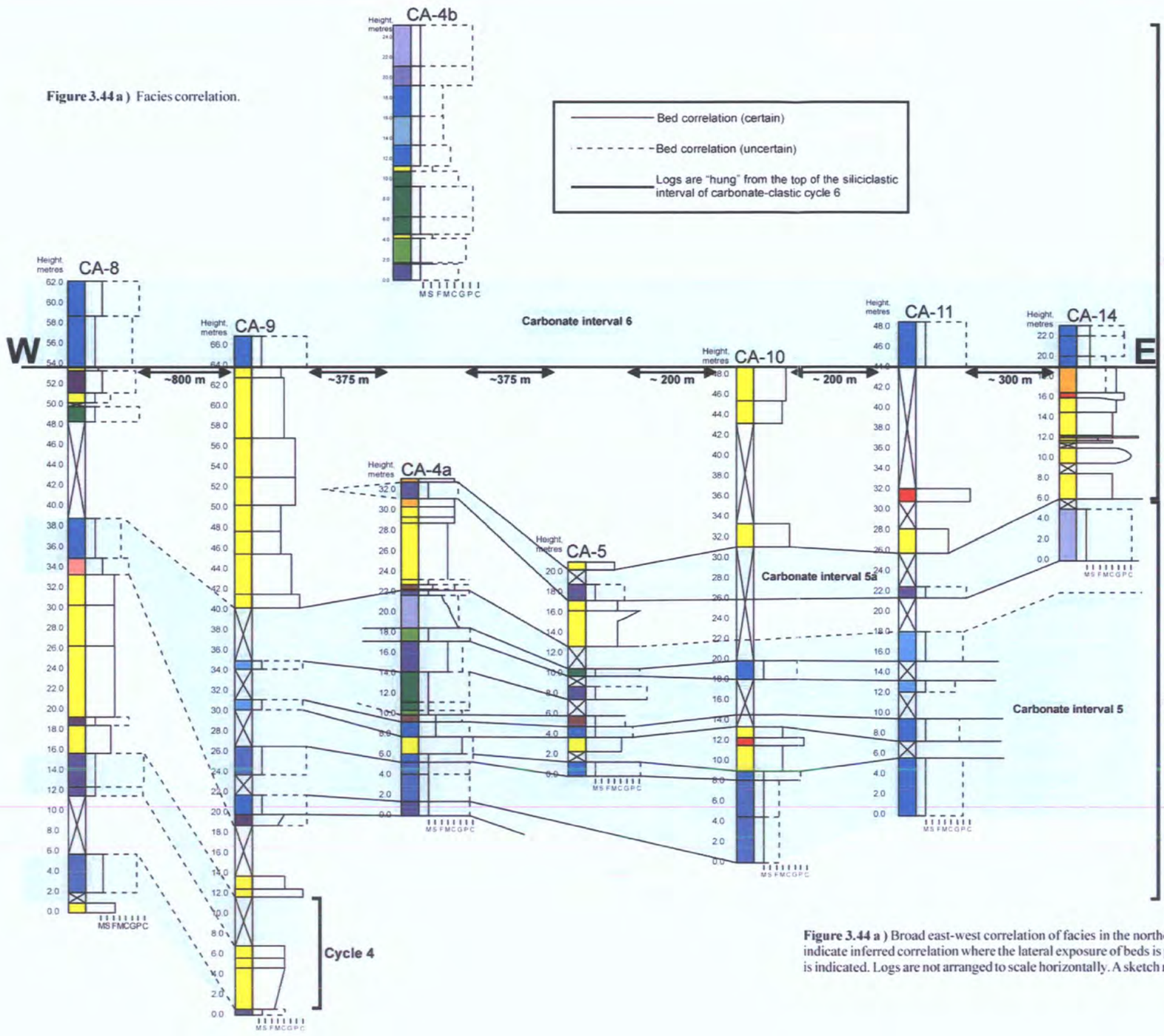
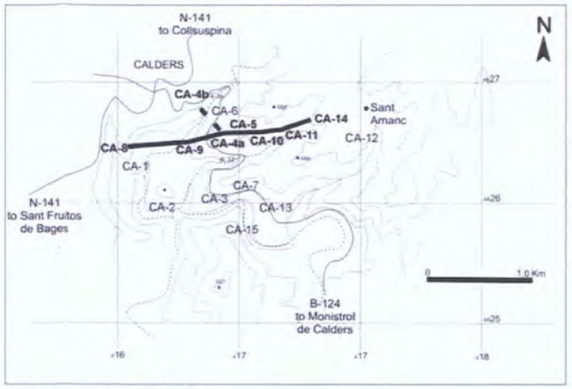


Figure 3.44 b) Sketch location map



Facies key

- Coral mixstone (**Section 3.3.1.1**)
- Coral domestone (**Section 3.3.1.2**)
- Coral bioclastic pack/rudstone (**Section 3.3.1.3**)
- Coral wacke/floatstone(**Section 3.3.1.4**)
- Coralgal foraminifera rudstone (**Section 3.3.1.5**)
- Coralgal foraminifera float/rudstone (**Section 3.3.1.6**)
- Foralgal grainstone (**Section 3.3.1.7**)
- *Gypsinal* grainstone (**Section 3.3.1.8**)
- Miliolid coralline algae pack/grainstone (**Section 3.3.1.9**)
- *Nummulites* siliciclastic pack/grainstone (**Section 3.3.2.1**)
- *Nummulites Discocyclina* siliciclastic packstone (**Section 3.3.2.2**)
- *Operculina* mudstone (**Section 3.3.2.3**)
- *Nummulites* coralline algae siliciclastic pack/grainstone (**Section 3.3.2.4**)
- *Gypsina* calcarenite (**Section 3.3.3.1**)
- *Nummulites* arenite (**Section 3.3.3.2**)
- Bioclastic siltstone (**Section 3.3.3.3**)
- Cross stratified calcarenite (**Section 3.3.4.1**)
- Lenticular polymict conglomerate (**Section 3.3.4.2**)

Figure 3.44 a) Broad east-west correlation of facies in the northern part of the Calders study area. All slid tie-lines have been traced in the field. Dashed lines indicate inferred correlation where the lateral exposure of beds is poor. Carbonate intervals are shaded in pale blue for clarity. Each carbonate-siliciclastic cycle is indicated. Logs are not arranged to scale horizontally. A sketch map showing locations of logged sections is present in **b)**.

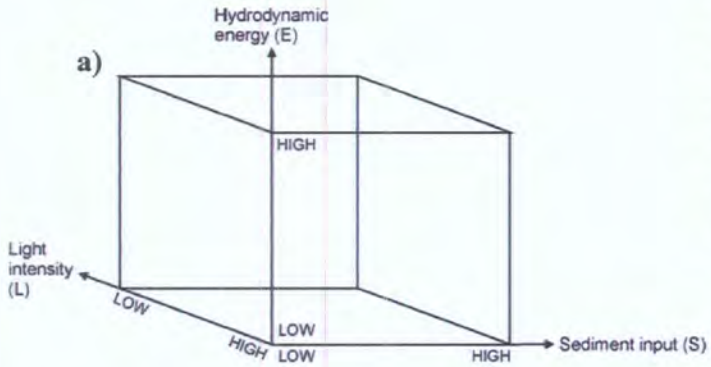
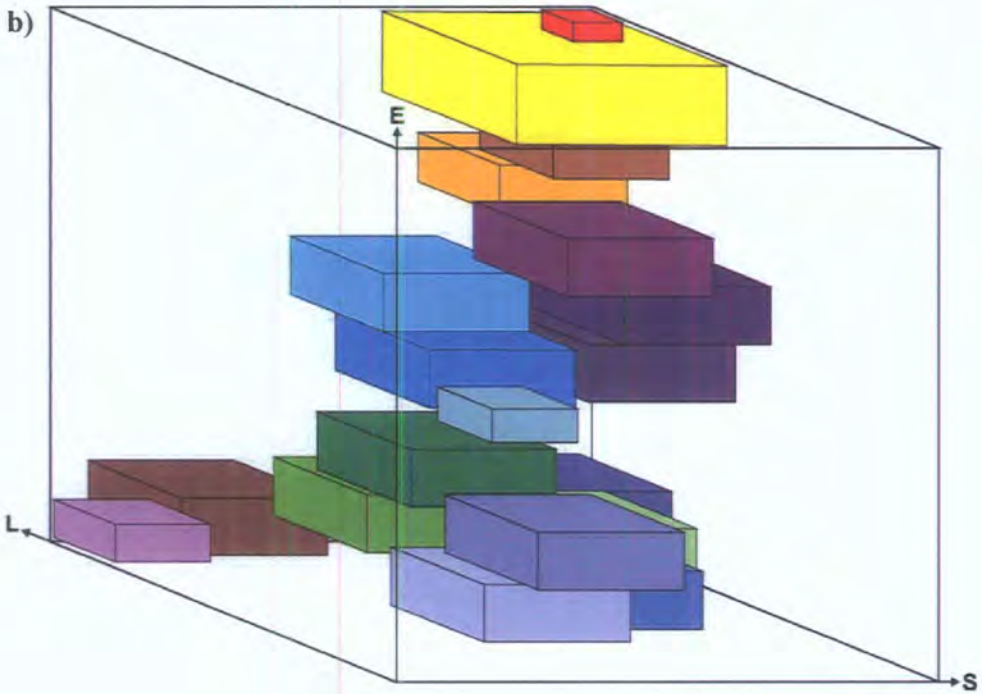


Figure 3.45a Environmental axes for a conceptual facies model modified from Insalaco *et al.* (1998). The volume defined by the three axes is referred to as LES space within which different facies will form. Light intensity (L) is inferred from palaeoecological criteria such as coral and larger benthic foraminifera growth forms. Hydrodynamic energy (E) is inferred from facies data i.e. presence/absence of matrix, clastic grainsize, fragmentation of bioclasts etc. Siliciclastic sediment input (S) is determined though calculation of weight percentage insoluble residue (see Chapter 6 and Appendix 1 for methodology). This diagram applies to environments with normal salinity, temperature and oxygenation.

Figure 3.45b Facies from the Calders succession plotted within LES space. The aim of this diagram is show the relative influence of light, energy and sediment input on different facies.



Facies key

- Coral mixstone (**Section 3.3.1.1**)
- Coral domestone (**Section 3.3.1.2**)
- Coral bioclastic pack/rudstone (**Section 3.3.1.3**)
- Coral wacke/floatstone(**Section 3.3.1.4**)
- Coralgal foraminifera rudstone (**Section 3.3.1.5**)
- Coralgal foraminifera float/rudstone (**Section 3.3.1.6**)
- Foralgal grainstone (**Section 3.3.1.7**)
- *Gypsinal* grainstone (**Section 3.3.1.8**)
- Miliclid coralline algae pack/grainstone (**Section 3.3.1.9**)
- *Nummulites* siliciclastic pack/grainstone (**Section 3.3.2.1**)
- *Nummulites Discocyclina* siliciclastic packstone (**Section 3.3.2.2**)
- *Operculina* mudstone (**Section 3.3.2.3**)
- *Nummulites* coralline algae siliciclastic pack/grainstone (**Section 3.3.2.4**)
- *Gypsina* calcarenite (**Section 3.3.3.1**)
- *Nummulites* arenite (**Section 3.3.3.2**)
- Bioclastic siltstone (**Section 3.3.3.3**)
- Cross stratified calcarenite (**Section 3.3.4.1**)
- Lenticular polymict conglomerate (**Section 3.3.4.2**)

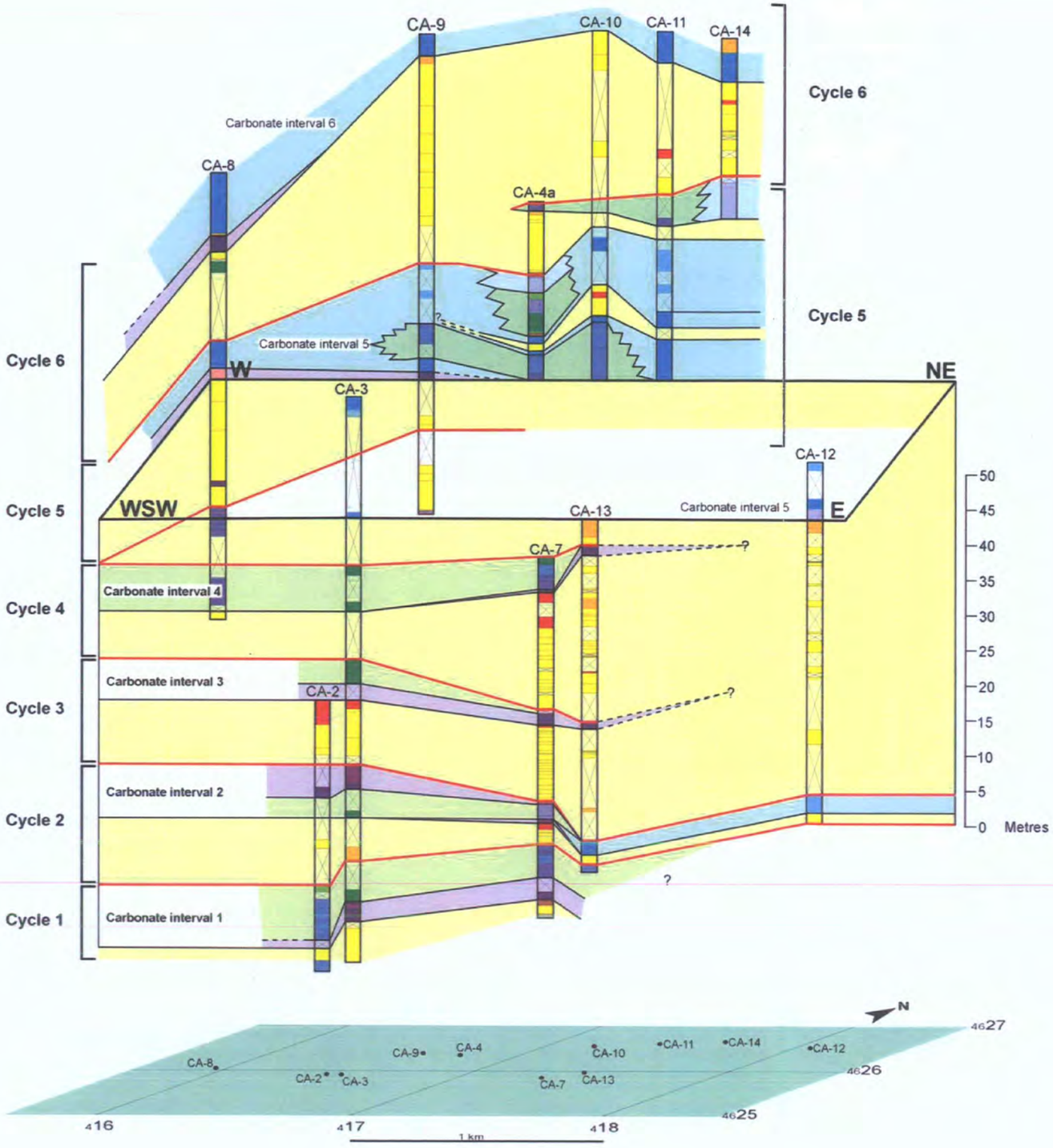


Figure 3.46 Area wide correlation of selected logs from the Calders study area illustrating the lateral continuity of facies associations. All logs are 'hung' from the base of carbonate interval 5 (thick black tie-line). Correlations have been traced in the field. Where tie-lines are dashed, correlations are inferred. Carbonate-siliciclastic cycles are illustrated. For clarity, the top of each cycle is marked by a red tie-line.

Facies associations

- Facies association 5
- Facies association 4
- Facies association 3
- Facies association 2
- Facies association 1

Facies key

- Coral mixstone (**Section 3.3.1.1**)
- Coral domestone (**Section 3.3.1.2**)
- Coral bioclastic pack/rudstone (**Section 3.3.1.3**)
- Coral wacke/floatstone(**Section 3.3.1.4**)
- Coralgal foraminifera rudstone (**Section 3.3.1.5**)
- Coralgal foraminifera float/rudstone (**Section 3.3.1.6**)
- Foralgal grainstone (**Section 3.3.1.7**)
- *Gypsina* grainstone (**Section 3.3.1.8**)
- Miliolid coralline algae pack/grainstone (**Section 3.3.1.9**)
- *Nummulites* siliciclastic pack/grainstone (**Section 3.3.2.1**)
- *Nummulites Discocyclina* siliciclastic packstone (**Section 3.3.2.2**)
- *Operculina* mudstone (**Section 3.3.2.3**)
- *Nummulites* coralline algae siliciclastic pack/grainstone (**Section 3.3.2.4**)
- *Gypsina* calcarenite (**Section 3.3.3.1**)
- *Nummulites* arenite (**Section 3.3.3.2**)
- Bioclastic siltstone (**Section 3.3.3.3**)
- Cross stratified calcarenite (**Section 3.3.4.1**)
- Lenticular polymict conglomerate (**Section 3.3.4.2**)

Table 3.3 Calders facies associations

Facies association	Facies	Dominant biota	Characteristics	Interpretation
Facies association 1 **(Section 3.4.1)**	Cross-stratified calcarenite, Lenticular polymict conglomerate, *Nummulites* litharenite, *Gypsina* calcarenite, *Operculina* mudstone	Rare biota include *Gypsina*, *Operculina*, miliolids, *Nummulites and* coralline algae.	Metre-scale cross-stratified calcarenites interbedded with silts, mudstones and foraminifera-rich sandstones.	**Shallow marine prograding siliciclastic shelf** Progradation of dunes with accumulation of fines during low-energy periods and within inter-dune areas Short-lived conglomeratic channels Colonisation of abandoned dunes by foraminifera
Facies association 2 **(Section 3.4.2)**	*Nummulites* siliciclastic pack/grainstone, *Nummulites Discocyclina* siliciclastic packstone, Nummulites coralline algae siliciclastic packstone, *Operculina* mudstone	Larger benthic foraminifera (*Nummulites, Discocyclina, Gypsina*) and coralline algae (encrusting forms and rhodoliths)	Well-bedded mixed carbonate-siliciclastic lithologies rich in larger benthic foraminifera. Crude cross-stratification and chaotic stacking of foraminifera tests.	**Current re-worked larger foraminifera and coralline algae accumulations** Colonisation of high-energy, unstable siliciclastic substrates No preserved topography
Facies association 3 **(Section 3.4.3)**	Coral bioclastic pack/rudstone, Coralgal foraminifera rudstone, Coralgal foraminifera float/rudstone, Coral wacke/floatstone	Corals, encrusting coralline algae and foraminifera (*Haddonia, Fabiania, Chapmanina*) with free-living and epiphytic foraminifera	Bedded units containing sparsely distributed *in situ* corals situated within a bioclastic wackestone to rudstone matrix.	**Sparse coral development** Development of coeval high and low-energy areas Constratal coral growth in muddy, siliciclastic-rich sediments No framework development or topography

Table 3.3 (cont.) Calders facies associations

Facies association	Facies	Dominant biota	Characteristics	Interpretation
Facies association 4 (**Section 3.4.4**)	Foralgal pack/grainstone, *Gypsina* grainstone, Miliolid coralline algae pack/grainstone	Foraminifera include *Gypsina, Calcarina, Rotorbinella, Amphitstegina, Nummulites,* miliolids and larger encrusting forms. Coralline algae present as laminar crusts, articulated forms and rhodoliths.	Intercalcated packstones and grainstones rich in abraded benthic foraminifera. Foraminifera dominated by *Gypsina.* Patchy coral growth.	**High-energy, winnowed shoals** Patchy coral development Deposition of fines during low-energy periods Possible sea floor vegetation upon which epiphytic foraminifera and articulated coralline algae were attached
Facies association 5 (**Section 3.4.5**)	Coral mixstone, Coral domestone	Corals. Various morphotypes present include branching, domal, platy and foliaceous.	*In situ,* densely packed coral colonies with a bioclastic packstone matrix situated in pockets between colonies.	**In situ coral growth** Localised framework development Supratstratal coral growth with a flourishing cryptic community

Individual facies development in the Calders area was influenced by a complex interaction of abiotic factors. The most important were light intensity, hydrodynamic energy levels and siliciclastic sediment input (**Figure 3.45**). The facies described in **Section 3.3** occur in 5 facies associations, summarised within **Table 3.3.** An area-wide correlation illustrating the lateral and vertical extent of facies associations is presented on **Figure 3.45**.

3.4.1 Facies association 1

Facies: Cross-stratified calcarenite
Lenticular polymict conglomerate
Gypsina calcarenite
Nummulites arenite
Operculina mudstone

Occurence

Facies association 1 comprises the majority of the studied succession at Calders (**Figure 3.46**). In addition, this was the first facies association to be deposited in the area, indicated by cross-stratified calcarenites overlying the prodelta Vespella Marls (log CA-7, beds 1 and 2). This association extends laterally > 2 km and beyond the limits of the study area. Facies association 1 attains a maximum thickness of 40 m and a minimum thickness of 2 m within cycle 1 (log CA-12, **Figure 3.46**). Typical thickness is around 10 m. Facies association 2 often succeeds this facies association vertically (**Figure 3.46**).

Characteristics

This facies association comprises sandstones, siltstones and shales, interbedded with foraminifera-rich sandstones and mudstones. **Figure 3.47** is a summary log of a typical succession through this facies association.

The majority of this association is composed of the cross-stratified calcarenite facies (**Figure 3.47**). Cross-stratified calcarenites occur intercalated with the *Gypsina* calcarenite (log 12, bed 15; log CA-13 bed 16), or the lenticular polymict conglomerate facies (log CA-7, bed 13; log CA-11, bed 7). Bed morphologies range from tabular to sigmoidal (**Figure 3.38**). Beds are cross-stratified, with set height up

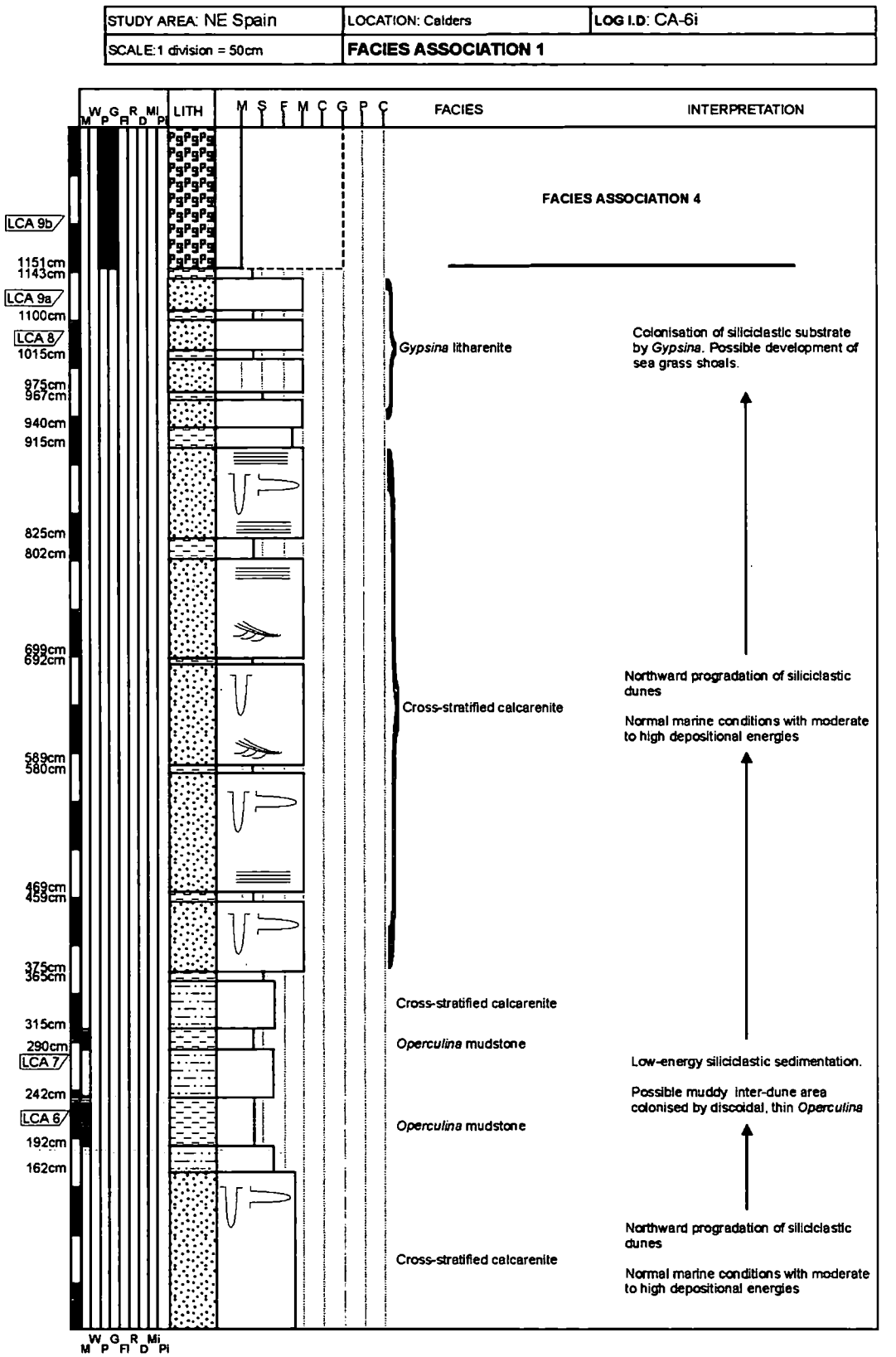


Figure 3.47 Summary log through a representative succession of facies association 1 (from log CA-6 part i). Facies association 4 in the upper portion of the log is situated at the base of carbonate interval 6. See Appendix 1 for key to log.

to 5 m (**Figure 3.38a**). Swaley cross-stratification with symmetrical ripple lamination superimposed on upper surfaces is evident within cycle 2 (**Figure 3.39a**). Palaeoflow directions towards the north and northeast are interpreted from assymetrical current ripple lamination. Sediments are largely unfossiliferous, although rare fragmented and abraded miliolids, coralline algae and small benthic foraminifera occur. Infaunal echinoids and large *Cerithium* gastropods are abundant on discrete horizons (**Section 3.3.4.1**). This facies contains vertical burrows, although burrows are terminated at bedding contacts.

The *Gypsina* calcarenite facies extends laterally for at least 1 km (**Figure 3.43**). Contacts with the cross-stratified calcarenite facies are sharp and non-erosional. *Gypsina* dominates the biota. Subordinate biota includes *Rotorbinella, Nummulites, Calcarina,* miliolids, echinoids, serpulids and bryozoa (**Section 3.3.3.1**). Most fossils in this facies are abraded (**Figure 3.35**). Siliciclastic grains are well sorted, subangular and subspherical. Mixed carbonate-siliciclastic mud-grade matrix is locally abundant. Winnowed intergranular areas are composed of equant calcite spar (**Figure 3.35a**).

Conglomerate units demonstrate lenticular morphologies up to 20 m in width and several hundred metres in length. Lateral equivalents of lenticular conglomeratic units are thin (<20cm) conglomeratic stringers intercalated calcarenite sediments (**Figure 3.48**). The lower contact between the lenticular polymict conglomerate and cross-stratified calcarenite facies is always erosional (**Figures 3.39a, 3.41b**). Scours are in-filled with massive, matrix-supported conglomerate (**Figure 3.41b**). The conglomerates are largely unfossiliferous, and may contain rare reworked *Nummulites* tests in some localities (log CA-7, bed 36).

The *Nummulites* arenite and *Operculina* mudstone facies are minor components of this facies association (**Figure 3.47**). The *Nummulites* arenite is encountered in cycle 1 (**Figures 3.43** and **3.46**). The lower contact with the cross-stratified calcarenite is gradational. The *Operculina* mudstone facies occurs intercalated with the cross-stratified calcarenite (**Figure 3.38c**). The *Operculina* mudstone facies is characterised by thin, often laminated mudstones and wackestones containing well-preserved, large, thin, discoidal foraminifera tests (**Figures 3.31** and **3.44**).

Figure 3.48 Pebbly stringers within the *Gypsina* calcarenite facies, interpreted as the lateral equivalent of the lenticular polymict conglomerate facies (log CA-4a, bed 17). Scale bar=50 cm

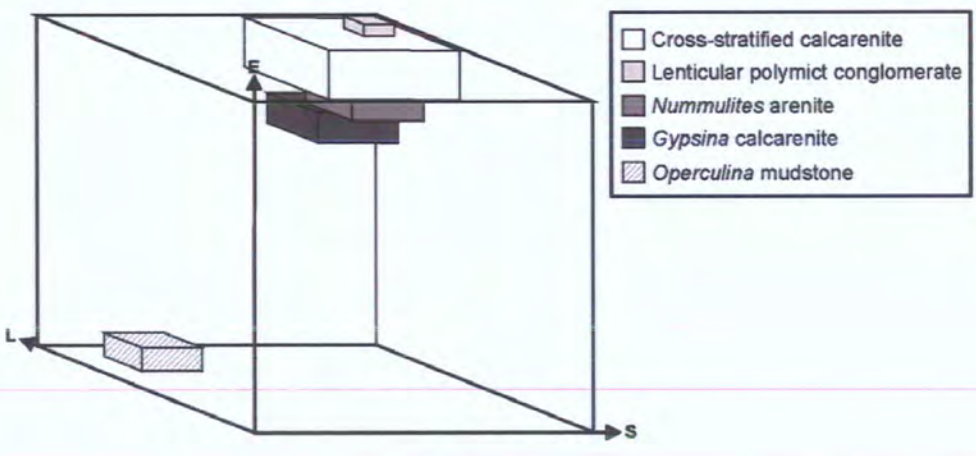


Figure 3.49 Placement of constituent facies of facies association 1 into LES space. This facies association is characterized by high siliciclastic sediment input and moderate to high depositional energies. The light intensity will vary according to grainsize of the sediment input i.e. fine vs. coarse sediment in suspension. Variable hydrodynamic energies are inferred from the presence of shale and silt interbeds. See text for discussion.

Interpretation

Normal open marine conditions are inferred for facies association 1 from the presence of stenohaline biota. Cross-stratification of beds and dune formation indicates a strong current influence and deposition above fair weather wave base (Reading 1996). Deposition between storm and fair weather wave base is inferred from the presence of swaley cross-stratified beds towards the base of the studied succession. It is interpreted that fine-grained interbeds and the *Operculina* mudstone facies accumulated during quiet periods and/or in inter-dune areas. Dunes are only exposed in two dimensions, thus the three-dimensional dune morphology cannot be determined.

It is postulated that substrate instability during deposition of the cross-stratified calcarenite facies inhibited colonisation by sessile carbonate-producing organisms. Mobile and infaunal organisms dominate the biota assemblage. Accumulation of the *Nummulites* arenite facies is interpreted to represent initial colonisation of the substrate as a consequence of reduced siliciclastic input and substrate stabilisation.

It is interpreted that the transition from cross-stratified calcarenites to *Gypsina* calcarenites represents stabilisation of the siliciclastic substrate and colonisation by sea grass and foraminifera. This was associated with a decrease in siliciclastic input, although water energies were high enough to rework siliciclastic grains and abrade biota (**Section 3.3.3.1**). The presence of sea grass in inferred from the abundance of *Gypsina* that often adapts (although not exclusively) to an epiphytic lifestyle in high-energy settings (Chaproniere 1975, Ghose 1977, Hohenegger *et al.* 1999, Walker *et al.* 2001).

Erosive-based lenticular conglomeratic units are interpreted as short-lived high-energy subaqueous marine channels (**Section 3.3.4.2**). There is no evidence for subaerial exposure of channels and development of interdistributary areas or palaeosol horizons. It is inferred that pebbly-stringers observed in laterally equivalent beds represent flooding events.

In summary, facies association 1 formed within a high-energy, shallow marine setting with high siliciclastic input (**Figure 3.49**). This association represents deposition and basinward (northward) progradation of large subaqueous dunes, with

colonisation by foraminifera as siliciclastic input decreased. The calcarenite facies association is part of the Centelles Formation that has been interpreted as deltaic in origin (see **Chapter 2**). There is not enough exposure in the Calders area to validate a deltaic origin.

3.4.2 Facies association 2

Facies: *Nummulites* siliciclastic pack/grainstone
Nummulites Discocyclina siliciclastic packstone
Nummulites coralline algae siliciclastic packstone
Operculina mudstone

Occurrence

Facies association 2 occurs throughout the studied succession, and is identified at the base of carbonate intervals 1 to 6 (**Figure 3.46**). The measured thickness of this association ranges from 1 to 3.5 m, and extends laterally > 1 km (**Figure 3.46**). This facies association typically succeeds the calcarenitic facies association. The lower contact between facies association 2 and lenticular polymict conglomerates of facies association 1 is gradational (**Figure 3.50**). In the absence of conglomerates, the transitional units are *Nummulites*-poor sands (log CA-7, beds 3 and 4). Contacts are sharp but non-erosional.

Characteristics

This facies association contains mixed carbonate and siliciclastic lithologies dominated by siliciclastic-rich larger benthic foraminifera and coralline algae dominated packstones and grainstones. Up to 50 wt. % of sediments are composed of non-carbonate material. **Figure 3.51** is a summary log of a typical succession through this association.

The *Nummulites* siliciclastic pack/grainstone is dominated by small robust and larger lenticular mono-specific *Nummulites* (**Figure 3.27c**) with a diverse foraminifera assemblage that includes *Gypsina, Amphistegina, Discocyclina, Calcarina,* miliolids and texturaliids (**Section 3.3.2.1**). Siliciclastic grain size ranges from clay to coarse-sand. Beds display crude cross-stratification (**Figure 3.27b**) and foraminifera chambers may be aligned in burrows.

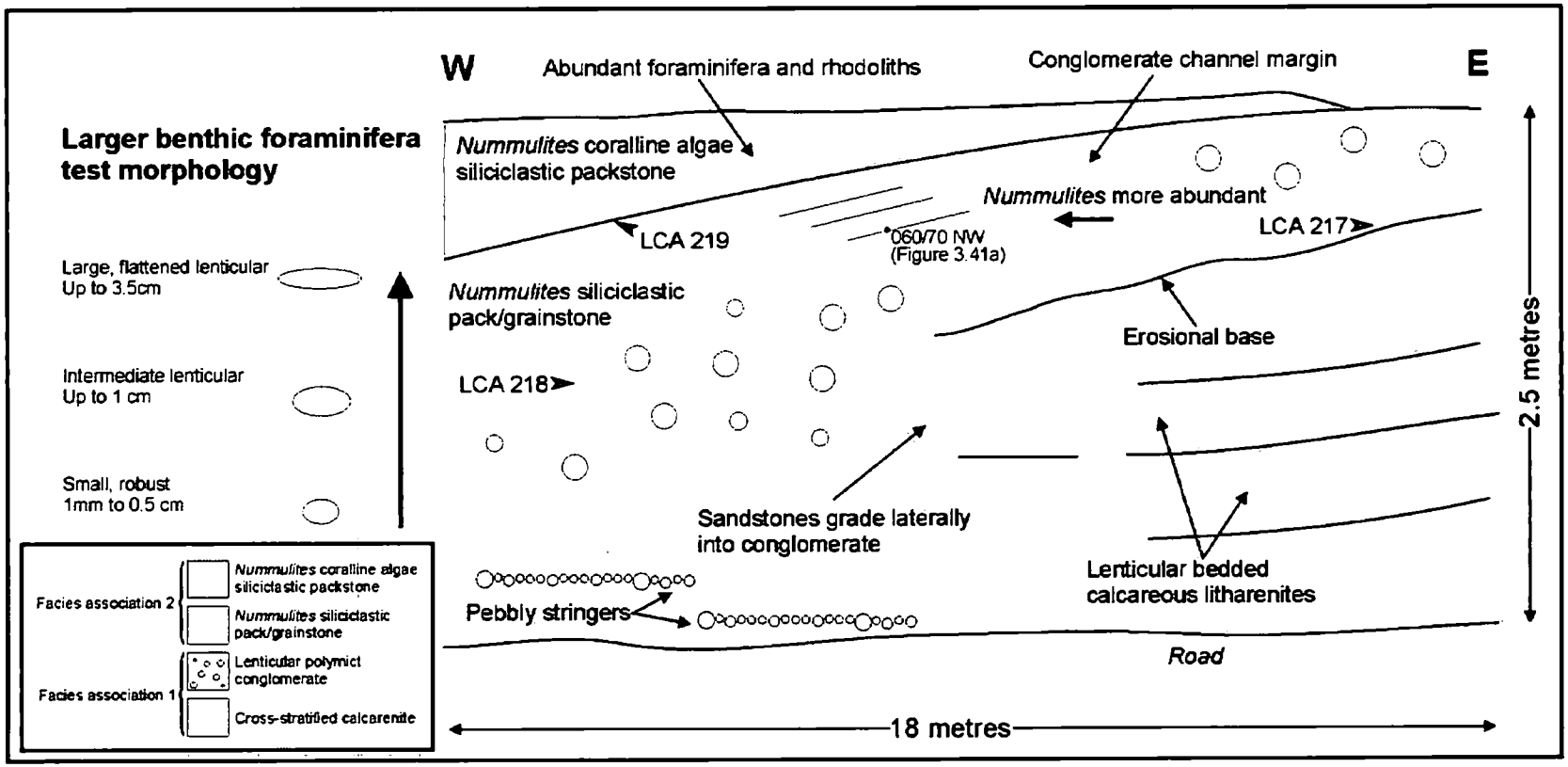


Figure 3.50 Field sketch of the contact between facies associations 1 and 2 (GR 16952614, facing N). Small robust *Nummulites* tests occur reworked within the underlying conglomerates. There is a noticeable vertical increase in foraminifera size (indicated) attributed to increasing energy levels and preferential winnowing of smaller tests by currents, or delayed reproduction and prolonged growth of foraminifera. See text for discussion.

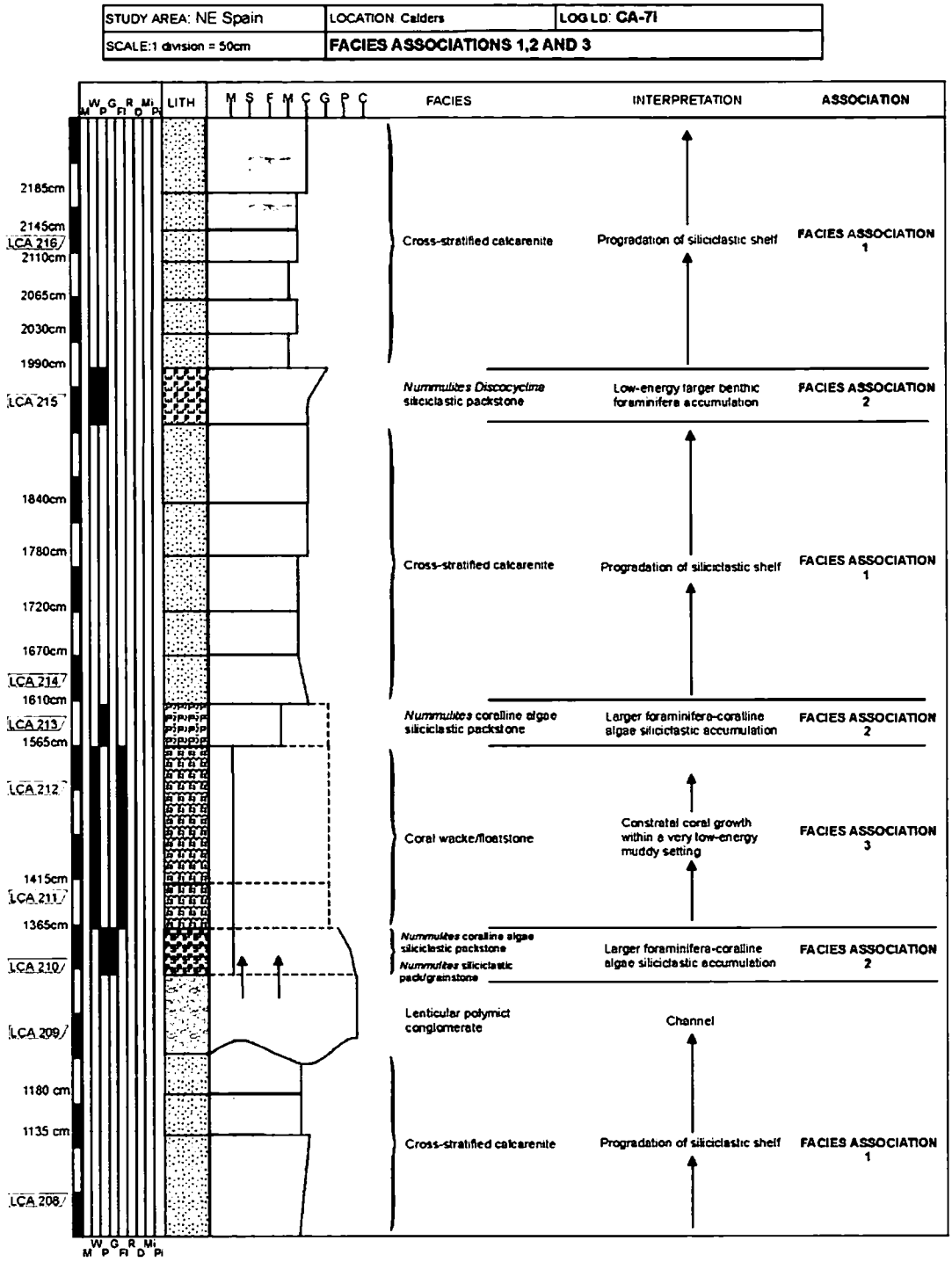


Figure 3.51 Summary log and interpretation through a portion of log CA-7 (part i) illustrating the stratigraphic relationship between the facies associations 1, 2 and 3. The lower contact of facies association 2 with facies association 1 is gradational. Within facies association 2, the *Nummulites* siliciclastic pack/grainstone facies is typically succeeded by the *Nummulites* coralline algae siliciclastic packstone. The *Nummulites Discocyclina* siliciclastic packstone may succeed this facies, but often occurs isolated within the calcarenites. See appendix 1 for key to log.

The *Nummulites* siliciclastic pack/grainstone is typically succeeded by the *Nummulites* coralline algae siliciclastic packstone facies. The contact is gradational over 5 to 10 cm. Sediments contain the same foraminifera assemblage as the preceding facies, although coralline algae (as rhodoliths and laminar crusts) becomes increasingly abundant and *Nummulites* are dominated by large robust lenticular forms up to 3.5 cm in diameter.

The *Nummulites Discocyclina* packstone occurs in the absence of the *Nummulites* coralline algae siliciclastic packstone facies. The contact between the two facies is gradational. Sediments are dominated by large (up to 2 cm), flattened, discoidal *Discocyclina* tests with lenticular intermediate *Nummulites* and subordinate *Operculina* (**Figure 3.30a**). Foraminifera tests display chaotic stacking (**Figure 3.30b**). In the western limit of the study area, this facies association is represented by accumulation of the *Operculina* mudstone facies (log CA-8, bed 10). Thin, discoidal *Operculina* tests occur intact within laminated muds.

Interpretation

This facies association contains a marine faunal assemblage. It is inferred from the stenohaline biota that open marine conditions prevailed throughout accumulation of this association (Ghose 1977, Hallock and Glenn 1986, Geel 2000). Deposition within the photic zone is inferred from the abundance of coralline algae and symbiont-bearing foraminfera (Chapronierre 1975, Ghose 1977, Hallock 1988). High-energy conditions are inferred from the presence of rhodoliths and thick, warty coralline algae crusts (Reid and MacIntyre 1988, Minnery *et al.* 1985, Minnery 1990). Reworking of sediments is inferred from the presence of cross-stratification and chaotic stacking of foraminifera tests (Aigner 1983, Racey 1988, 2001). This occurred *in situ* as the state of preservation of tests indicates that have not undergone transportation. This *in situ* reworking has locally removed fines and fragmented more fragile bioclasts such as bryozoa and molluscs.

The abundance of non-carbonate material in the matrix of sediments indicates their accumulation under the influence of a siliciclastic input and/or the influence of currents reworking pre-deposited siliciclastic sediments. It is inferred that unstable substrates inhibited the development of a sessile calcareous organisms such as corals. Larger benthic foraminifera such as *Nummulites* were able to use their pseudopodia to

extract themselves if buried (van der Zwann *et al.* 1999). Progressively more stable environmental conditions with a decrease in coarse siliciclastic sediment input are inferred from the progressive vertical increase in the abundance of coralline algae, although maintained high-energies are inferred from the dominance of larger *Nummulites* tests over smaller forms that have been preferentially winnowed (cf. Aigner 1982, 1983).

Large, flattened *Discocyclina* and *Operculina* tests are characteristic of poorly illuminated conditions (Hottinger 1983, Leutenegger 1984, Hallock and Glenn 1985, Trevisani and Papazzoni 1996). It is postulated that *Nummulites Discocyclina* siliciclastic packstone and *Operculina* mudstone facies were deposited in relatively deep, protected or turbid waters. Siliciclastic sediments within the matrix are clay to silt grade, and would have remained in suspension before deposition.

In summary, facies association 2 represents reworked larger foraminifera and coralline algae accumulations that developed on siliciclastic substrates within the photic zone (**Figure 3.52**).

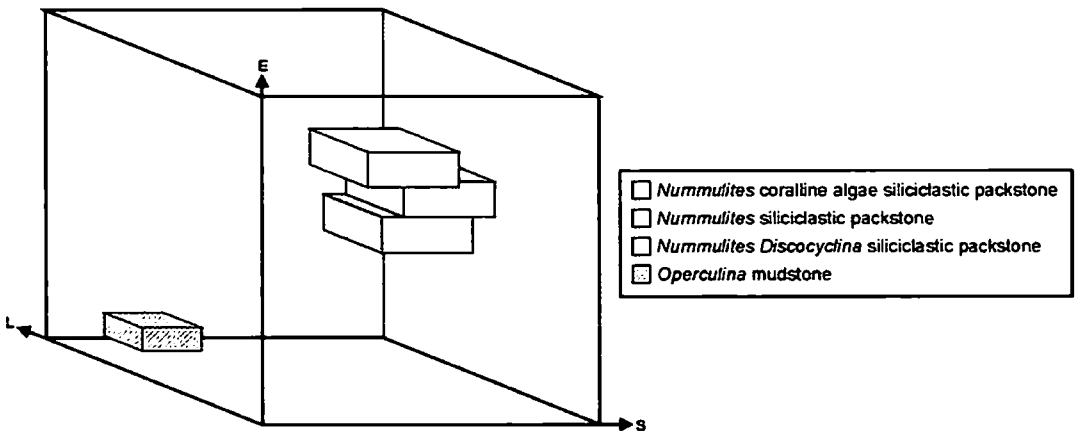


Figure 3.52 Placement of the constituent facies of facies association 2 into LES space. Most facies were deposited within moderate to high-energy settings characterised by high siliciclastic input and variable light intensities. The *Operculina* mudstone facies formed within poorly illuminated conditions. Although the matrix contains siliciclastic material, low energy conditions and the fine-grained nature of sediments resulted in low rates sediment of sediment accumulation.

3.4.3 Facies association 3

Facies: Coral bioclastic pack/rudstone
Coralgal foraminifera rudstone
Coralgal foraminifera float/rudstone
Coral wacke/floatstone

Occurrence

Facies association 3 is encountered in all carbonate intervals of the Calders succession (**Figure 3.46**). This facies association is up to 9 m in thickness and extends laterally > 1 km (**Figure 3.46**). This association is often intercalated with facies association 4 (**Figure 3.46,** carbonate interval 5), and overlies facies association 2 (**Figure 3.53**). The contact is sharp but non-erosional. In some localities, the lower contact with facies association 2 is often enhanced by later dissolution during burial (**Figure 3.29c**).

Characteristics

This facies association contains carbonate-dominated sediments with a variable abundance of corals, coralline algae, encrusting foraminifera and fine-grained matrix. Textures include wacke/floatstone, rudstone, float/rudstone and pack/rudstone. **Figure 3.53** is a typical log through this facies association.

The first facies to be deposited is typically the coral wacke/floatstone. Platy corals are locally abundant where this facies overlies the larger benthic foraminifera coralline algae facies association (**Figure 3.29c**). Sediments are composed of sparsely distributed small branching corals situated within a muddy mixed carbonate-siliciclastic matrix (**Figure 3.17**). Matrix can comprise up to 80 % of sediments, and clay to silt-grade siliciclastic material comprises 12 to 34 wt. %. Coral colonies may demonstrate thin coralline algae crusts (**Figure 3.17c**) with fragile bioclasts such as bryozoa well preserved in the matrix (**Figure 3.17d**).

The coral wacke/floatstones are succeeded by and/or intercalated with coral bioclastic pack/rudstones (**Figure 3.53**). The contact is sharp but non-erosional. Sediments are characterised by sparsely distributed *in situ* coral colonies that are often overturned (20 to 30 % of facies volume) (**Figure 3.15a**). The intervening pack/rudstones contain abundant coral, coralline algae and foraminifera debris (**Figures 3.15b** and **c**). Fragmented bioclasts show minor abrasion. Foraminifera

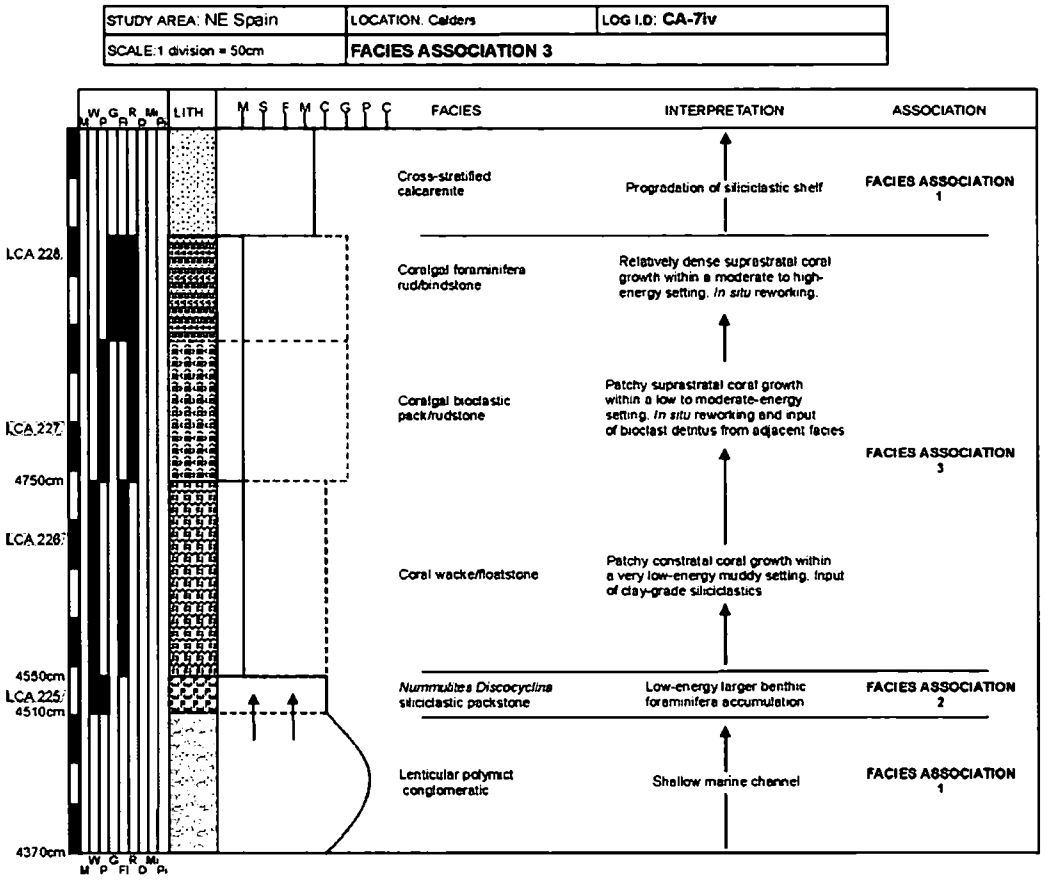


Figure 3.53 Typical log through facies association 3 (log CA-7 part iv, carbonate interval 4). This association typically overlies facies association 2 and is overlain by facies association 1. In this instance, facies association 3 forms an apparent shallowing-upward succession. See appendix 1 for key to log.

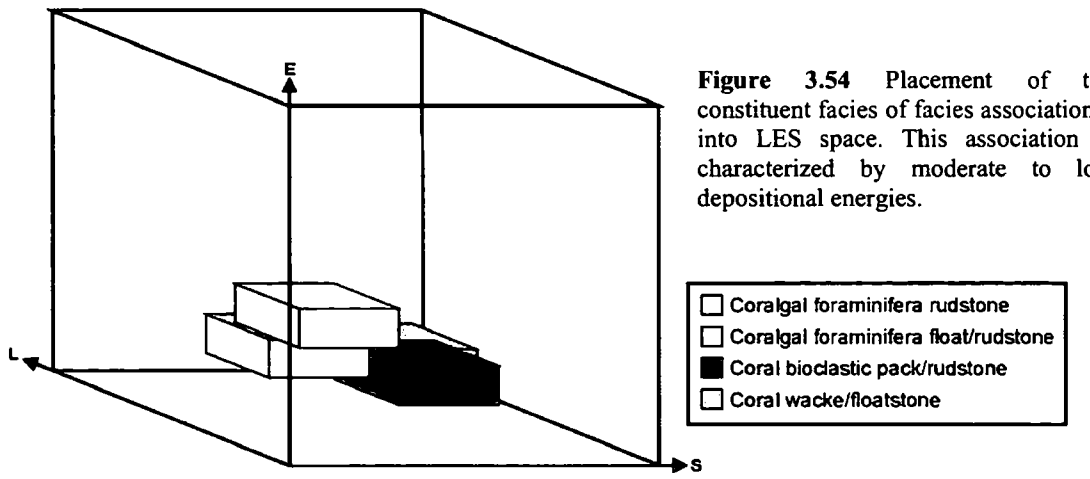


Figure 3.54 Placement of the constituent facies of facies association 3 into LES space. This association is characterized by moderate to low depositional energies.

include benthic forms (*Amphistegina and Gypsina*), with large encrusting *Haddonia, Fabiania, Chapmanina* and victoriellids (**Figure 3.16b**). Concentric laminar rhodoliths with delicate and columnar branches up to 7.5 cm in diameter are also present. Clay to silt-grade siliciclastics comprise 7.5 to 20.5 wt. % of sediments (**Section 3.3.1.3**).

The final sediments to be deposited in this association are coralgal foraminfera rudstones and float/rudstones (**Figure 3.53**). These sediments are characterised by *in situ* branching, domal and foliaceous coral colonies with an intervening bioclastic matrix. Corals often demonstrate thick laminar and warty coralline algae-foraminifera crusts (**Figures 3.18a, 3.19a** and **3.20b**). Spheroidal concentrically laminated rhodoliths are common (**Figure 3.20c**). Sediments contain a diverse foraminifera assemblage dominated by encrusting forms (*Gypsina, Haddonia, Fabiania*) with miliolids, *Amphistegina* and small hyaline benthics. Bioclasts are situated within a mixture of micrite (up to 65 %) and clay to silt grade siliciclastic material (5 to 21 wt. %).

Interpretation

This facies association contains a marine faunal assemblage, with normal open marine conditions indicated from the stenohaline biota. The coral wacke/floatstones are interpreted to have formed within a low-energy setting with a significant input of fine-grained siliciclastic sediment that would have been deposited from suspension (**Section 3.3.1.4**). It is interpreted that the initial phase of platy coral growth on the larger foraminifera coralline algae facies association is attributed to low light levels with fine-grained siliciclastic sediments kept in suspension. A progressive increase in the relative abundance of delicate branching corals is attributed to increased deposition of suspended particulate sediments through decreasing water energy and/or increased water clarity through reduced suspended sediment input. It is inferred from the negligible cryptic community that branching corals were living constratally (**Section 3.3.1.4**).

The coral bioclastic pack/rudstones, coralgal foraminifera rudstone and coralgal foraminfera float/rudstone facies contain a greater abundance and diversity of *in situ* corals. It is inferred from the flourishing cryptic community (crustose coralline algae and large encrusting foraminfera) that corals were growing suprastratally,

although there is no evidence to suggest the development of any topographic relief (**Sections 3.3.1.3, 3.3.1.5** and **3.3.1.6**). Moderate to high-energy conditions and competition for space is inferred from the development of thick, warty coralline algae crusts, particular within rudstones (Reid and MacIntyre 1988, Minnery *et al.* 1985, Minnery 1990). Agitated conditions are confirmed from the presence of rhodoliths (Bosence 1983b). *In situ* fragmentation and limited transportation of organisms is inferred from moderate to low abrasion of bioclasts.

A moderate siliciclastic input is indicated from the significant percentage of non-carbonate material. Intermittent low-energy periods are inferred from the fine grainsize of siliciclastics and the local abundance of micrite. Additionally, the suprastratal coral growth fabrics can locally modify the hydrodynamic conditions, creating sheltered areas and trapping suspended sediment (Insalaco 1998).

In summary, it is interpreted that facies association 3 represents patchy *in situ* coral development within a shallow marine shelf setting. Rudstone textures with abundant encrusting organisms formed in local high-energy areas. Float/rudstones and pack/rudstones formed in adjacent low-energy areas receiving the detritus of erosion of the local coral accumulations. Coral wacke/floatstones formed within the lowest-energy setting within neighbouring shallow, muddy partially protected environments (**Figure 3.54**).

3.4.4 Facies association 4

Facies: Foralgal pack/grainstone
Gypsina grainstone
Miliolid coralline algae pack/grainstone

Occurrence

This facies association is encountered in carbonate intervals 2, 5 and 6 (**Figure 3.46**). Within carbonate interval 6, the foraminifera grainstone facies association can be traced laterally for over 2 km (**Figure 3.46**). The measured thickness of sediments ranges from 1 to 3.5 m, although within carbonate interval 5 this association is up 20 m thick (**Figure 3.46**). This facies association often succeeds facies association 1 where the larger benthic foraminifera and coralline algae association is absent or very thin (e.g. carbonate interval 6, **Figure 3.46**). The contact with facies association 1 is sharp and non-erosional. This facies also occurs intercalated with facies association 3

(i.e. log CA-4a, beds 4 to 12), and may be succeeded by facies association 5 (**Figure 3.46**).

Characteristics

This facies association is composed of carbonate-dominated sediments with packstone and grainstone textures. **Figure 3.55** represents a typical log through this succession. If present, the miliolid coralline algae pack/grainstone is the first facies to be deposited in this association. The contact between the miliolid coralline algae pack/grainstone and overlying foralgal pack/grainstone facies is gradational. Sediments are moderately well sorted. Abraded miliolids and coralline algae fragments with subordinate *Gypsina* and *Orbitolites* are the dominant biota. Clay to silt grade siliciclastic material comprises up to 25.5 wt. % of sediments.

The majority of this facies association is composed of intercalated *Gypsina* grainstones and foralgal pack/grainstones (**Figure 3.53**). The two facies differ only in that *Gypsina* dominates the former, comprising up to 25 % of sediments. Other foraminifera present in abundance are miliolids, texturaliids, *Rotorbinella, Calcarina, Amphistegina, Nummulites, Haddonia, Chapmanina, Fabiania* and victoriellids. Coralline algae occur as spheroidal concentric laminar rhodoliths, peloids and articulated forms. Patchily distributed i*n situ* branching corals are observed, although corals do not make up a framework. Intergranular areas are largely cemented with calcite spar (up to 59 %), although micrite and clay to silt-grade siliciclastic material can comprise up to 20 % of sediments.

Interpretation

Facies association 4 contains a marine faunal assemblage and deposition within the photic zone is inferred from the presence of symbiont-bearing foraminifera such as *Amphistegina, Calcarina* and *Nummulites* (Ghose 1977, Hallock and Glenn 1985, Hohenegger *et al.* 1999). Moderate to high-energy conditions are inferred from the pack/grainstone textures, abraded bioclasts and the dominance of robust *Nummulites* and *Amphistegina* with thick test walls (Hallock and Glenn 1985, Hallock 1988). Intermittent, low-energy periods are inferred from the accumulation of clay to silt grade siliciclastic material. Accumulation of fines may have been locally enhanced through the presence of *in situ* corals (see **Sections 3.3.1.7** and **3.3.1.8**). It is

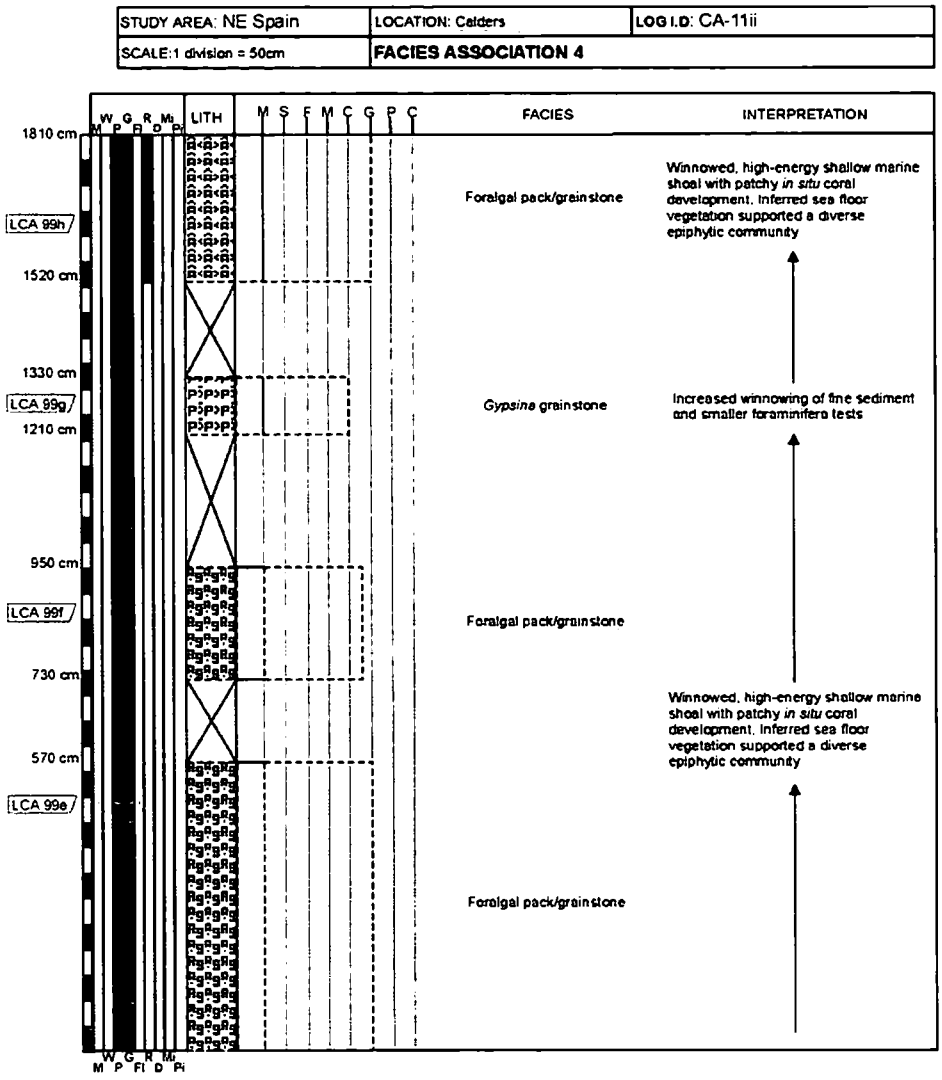


Figure 3.55 Typical logged section through facies association 4 (log CA-11, carbonate interval 5). The miliolid coralline algae pack/grainstone facies (not present in this section) is observed only within carbonate intervals 5 and 6. See appendix 1 for key to log.

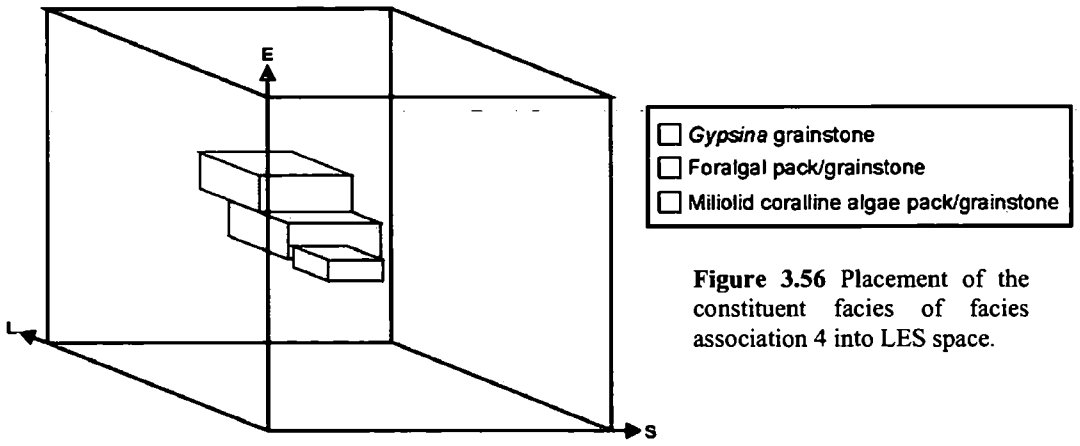


Figure 3.56 Placement of the constituent facies of facies association 4 into LES space.

postulated that sea grass (not preserved) was abundant in this environment. Miliolids, *Amphistegina*, *Calcarina* and articulated coralline algae can adapt to an epiphytic mode of life in high-energy conditions where substrates may be unstable or unsuitable (Ghose 1977, Carbone *et al.* 1994, Hohenegger *et al.* 1999). It is inferred from the abundance of miliolids and dearth of rotaliids that the miliolid coralline algae pack/grainstone facies was deposited in a shallower setting than the foralgal pack/grainstone and *Gypsina* grainstones.

In summary, facies association 4 formed within a moderate to high-energy marine environment within the upper parts of the photic zone (**Figure 3.56**). This environment was subject to fine-grained siliciclastic sediment input during low-energy periods. Accumualtion of fines was also encouraged through the baffling effect of corals and sea floor vegetation.

3.4.5 Facies association 5

Facies: Coral mixstone
Coral domestone

Occurrence

Facies association 5 is present in carbonate intervals 5, 5a and 6 (**Figure 3.46**). This facies association can be traced laterally 50 to 100 m, and measured thickness ranges from 3 to 10 m (**Figure 3.57**). This association is intercalated with facies association 3 (log CA-4a beds 12 and 13) and facies association 4 (log CA-4b beds 12 and 13). The upper and lower facies association contacts are sharp and non-erosional.

Description

This facies association is composed of carbonate-dominated sediments that have mixstone and domestone textures. **Figure 3.57** is a summary log of a typical succession through this facies association.

Where present, the coral domestone underlies the coral mixstone. The upper contact with the coral mixstone is gradational, marked by a gradual diversification in coral colony morphology over a distance of approximately 50 cm (log CA-4b, beds 12 and 13). Coral mixstone comprises most of this association (**Figure 3.57**). *In situ* corals dominate the biota, with coral skeletal volumes up 55 %. Colonies demonstrate branching, domal, foliaceous and massive morphologies (**Figures 3.10** and **3.13**), and

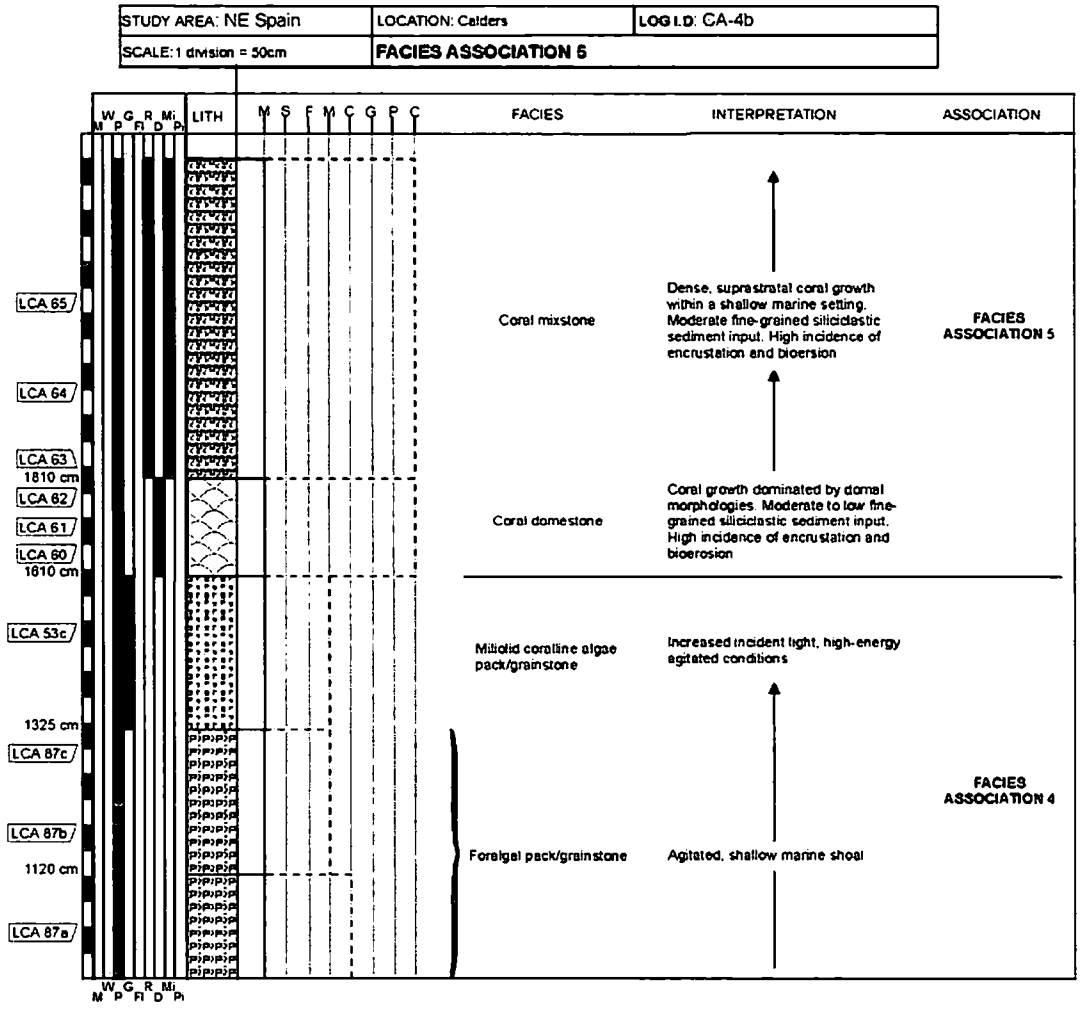


Figure 3.57 Typical logged section through facies association 5 (from log CA-4b). This association is commonly associated with facies association 4. See appendix 1 for key to log.

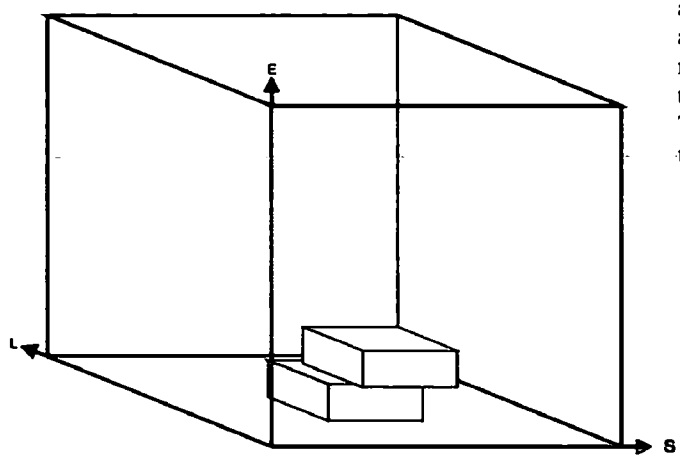
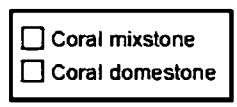


Figure 3.58 Placement of facies association 5 into LES space. This association formed within moderately low energy conditions in the upper parts of the photic zone. This facies association formed under the influence of a siliciclastic input.



locally make up a framework (although there is no evidence to suggest they formed significant topographic relief). Coralline algae and larger foraminifera may encrust corals. Corals have also been extensively bored by lithophagids (**Figure 3.11c** and **d**). A bioclastic packstone matrix is preserved within pockets between coral colonies. Up to 27.5 wt. % of the matrix is clay-grade siliciclastic material. Well-preserved, unabraded bioclasts in the matrix include laminar coralline algae, encrusting foraminifera (*Haddonia, Fabiania,* victoriellids and acervulinids), bryozoa and molluscs. Elliptical rhodoliths up to 5 cm in diameter show variable growth morphologies, from tight concentric laminae to delicate branches (**Figure 3.12**).

Interpretation

It is interpreted from the abundance of *in situ* corals that this facies association formed within the photic zone in an open marine setting. Suprastral growth and moderate nutrient influxes are inferred from the flourishing cryptic community and boring of coral colonies by lithophagids. Variable depositional energies are inferred from the changing growth morphology of rhodoliths. A siliciclastic input is inferred from the abundance of fine-grained siliciclastic sediment in the matrix.

In summary, facies association 5 is interpreted as localised, dense *in situ* coral development within a shallow marine setting periodically influenced by a fine-grained siliciclastic input (**Figure 3.58**).

3.4.6 Summary and depositional model

Each individual carbonate-siliciclastic cycle represents the deposition and colonisation of a siliciclastic substrate, initially by a foraminifera-dominated community and then diversification into a coralgal-dominated community. The following summarises the temporal and spatial evolution of facies in the Calders area. A palaeoenvironmental model, based on the conclusions in the following, is presented on **Figure 3.61**.

Cycle 1

The first carbonate-siliciclastic cycle was deposited on the arenaceous marls of the Vespella Formation. Up to 5 m of siliciclastic sediments were deposited (facies association 1). Siliciclastic sediments show a broad northward progradation direction.

A decrease in siliciclastic input and onset of the carbonate factory is represented by deposition of facies association 2. Up to 1.7 m of nummulitic and coralline algae-rich sediments were deposited (**Figure 3.56**), representing development of *Nummulites* and coralline algae siliciclastic shoals. A progressive increase in hydrodynamic energy selectively winnowed smaller foraminifera tests, with deposits containing abundant *Nummulites* up to 3.5 cm in diameter within the highest-energy settings. These deposits can be traced laterally >500 m (**Figure 3.46**).

The cessation of coarse-grained siliciclastic input and a decrease in hydrodynamic energy is interpreted from the deposition of facies association 3. Low-energy, partially protected conditions prevailed in the east (log CA-7), indicated by accumulation of 4.6 m of the coral wacke/floatstone facies. Contemporaneous moderate to high-energy conditions prevailed approximately 700 m to the west (logs CA-2 and CA-3), where the coralgal foraminifera rudstone and coral bioclastic pack/rudstone facies formed. It is interpreted that the area to the west was characterised by patchy coral development with localised (< 50 m in diameter) areas of more dense coral growth. It is postulated that the development of corals to the west provided protection from marine currents to the area in the east.

Cycle 2

Carbonate development ceased as a consequence of siliciclastic input. The sharp, non-erosional contact between facies associations 3 and 1 indicates that the change from carbonate development to siliciclastic sedimentation was rapid. Up to 6 m of siliciclastic sediments were deposited, thinning to 2.4 m towards the east (CA-13). High-energy, shallow marine conditions in the southeast are inferred from the presence of the lenticular polymict conglomerate facies (log CA-7).

Decreasing siliciclastic input and initiation of the carbonate factory is inferred from the deposition of facies association 2, although the lateral extent of sediments cannot be determined due to lack of continuous exposure. High hydrodynamic energy is inferred from the large *Nummulites* test size and abundance of rhodoliths.

The cessation of sand-grade siliciclastic input and the development of more stable substrates are interpreted from the development of facies association 3 (**Figure 3.46**). As described for cycle 2, low-energy conditions existed in the east with

the accumulation of the coral wacke/floatstone facies (log CA-7) with relatively high-energy conditions with abundant *in situ* corals towards the west (log CA-3). The development of a contemporaneous shallow, high-energy foralgal shoal setting approximately 200 m to the east is evident from the gradation of facies association 3 into facies association 4 between logs CA-7 and CA-13.

Termination of carbonate production within cycle 2 was less abrupt than cycle 1, evident by the deposition of up to 3.4 m of facies association 2 above the coral-dominated facies (**Figure 3.46**). The contact is gradational over 10 cm. It is interpreted that increasing rates of siliciclastic input caused mortality of sessile biota such as corals. However, siliciclastic input was within the tolerable limits of more mobile organisms such as *Nummulites* and rhodoliths.

Cycle 3

The mortality of most calcareous benthic organisms resulted from the area-wide deposition of up to 13 m of facies association 1 (**Figure 3.46**). High-energy conditions prevailed, inhibiting the development of a sessile benthic community. The dominant biota were infaunal burrowing echinoids and soft-bodied beasts, although a thin accumulation of facies association 2 is evident (log CA-7, bed 21). Larger foraminifera are dominated by thin, discoidal *Discocyclina*, thus were deposited under low-light conditions (Hallock 1988, Racey 1988, 2001). This is interpreted as a localised phenomenon as this interval cannot be traced laterally.

Widespread (> 1 km) deposition of up to 1.7 m of facies association 2 represents a decrease in siliciclastic input and onset of carbonate production. High-energy conditions are inferred from *Nummulites* tests reworked in conglomeratic channels (**Figure 3. 50**).

The development of thick carbonate units is evident in the west of the study area (log CA-3), with deposition of up to 2.4 m of facies association 3. It cannot be ascertained if carbonate development occurred during cycle 3 towards the east due to lack of exposure (**Figure 3.46**). Patchy, locally dense coral development (CA-3) was contemporaneous with the deposition of coral detritus (coral bioclastic bioclastic pack/rudstones) 500 m to the northwest (log CA-8).

Cycle 4

Widespread deposition of up to 13 m of facies association 1 halted carbonate development in Cycle 3. Facies association 1 thins to 10 m down-dip (towards the west and northwest) (**Figure 3.46**). Rapid sedimentation and burial of biota is inferred from the sharp lower contact with facies association 3. Subsequent accumulation of 0.4 to 1.5 m of facies association 2 indicates initiation of the carbonate factory as siliciclastic input decreased.

With continued decrease in siliciclastic input, carbonate-dominated deposition resumed. As described in the previous cycles, low-energy conditions with abundant clay-grade siliciclastic input prevailed in the east (CA-7), with accumulation of the coral wacke/floatstone facies. Carbonate development in the far eastern limit of the study area (logs CA-12 and CA-13) can only be inferred. Contemporaneous moderate-energy conditions with patchy, locally dense coral development prevailed approximately 150 m to the west (CA-3) with accumulation of the coralgal foraminifera rudstone facies. Low to moderate-energy conditions existed 1 km to the northwest, with accumulation of the coral bioclastic pack/rudstone facies (CA-8).

Cycle 5

Widespread deposition of facies association 1 terminated carbonate production of Cycle 4. Facies association 1 thins from 16 m in the west (CA-8) to 6 m 2 km to the east (**Figure 3.46**). The subsequent cessation of siliciclastic input in the east (CA-12) is marked by the deposition of the *Gypsina* calcarenite facies (**Figure 3.43** and **3.46**). In the central part of the study area (CA-9), the transition is marked by the *Nummulites* siliciclastic packstone facies (**Figures 3.44** and **3.46**). The transition is marked by the deposition of the *Operculina* mudstone facies to the west (**Figures 3.44** and **3.46**). An east to west decrease in hydrodynamic energy, siliciclastic input and possibly incident light is interpreted from this facies trend.

Two carbonate intervals developed during cycle 3. The main carbonate interval (interval 5) thickens significantly from 5 m in the west (CA-8) to approximately 18 m in the east (CA-11) (**Figure 3.46**). Carbonate interval 5 is dominated by intercalation of facies associations 3 and 4, and culminates with the development of facies association 5 (**Figure 4.46**). The development of contemporaneous high-energy, shallow foralgal shoals and relatively lower-energy coral development in adjacent areas is interpreted from this association. More

persistent high-energy conditions are inferred for the east from the thick accumulation of facies association 4 (CA-11) (**Figures 3.44** and **3.46**). Unstable environmental conditions and shifting facies belts are inferred from the intercalation of detrital facies with facies containing a significant proportion of *in situ* corals. Development towards stable environmental conditions is inferred from the culmination in a coral mixstone (**Figure 3.46**).

Carbonate interval 5a thickens from < 2 m to 6 m over a distance of approximately 1 km (**Figures 3.44** and **3.46**). Facies association 3, composed of coral wacke/floatstones) passes laterally into facies association 5 (**Figure 3.46**). A progressive increase in water energy from west to east is inferred from this arrangement of facies.

Cycle 6

As described in all of the previous cycles, widespread deposition of 12 to 20 m of facies association 1 terminated carbonate production in Cycle 5. Facies association 1 within cycle 6 culminates with the deposition of the *Gypsina* calcarenite facies (**Figure 3.47**). This facies demonstrates a progradational morphology and is observed to thin significantly down-dip towards the northeast (**Figure 3.59**). This facies represents the colonisation of an unstable siliciclastic substrate by sea floor vegetation and non-symbiont-bearing larger benthic foraminifera (**Sections 3.3.3.1** and **3.4.1**).

It is interpreted that the carbonate factory was established as siliciclastic input decreased and water became less turbid, allowing colonisation by symbiont-bearing biota (**Figure 3.59c**). Facies association 4 formed within high-energy conditions on abandoned foresets and thickened down-slope roughly northward (**Figure 3.59c**). The base of carbonate interval 6 is marked by the wide spread (>2 km) accumulation of up to 8 m of facies association 4 (**Figures 3.44** and **3.46**). Facies association 4 passes laterally (down-slope) into coral wacke/floatstones of facies association 3 (**Figure 3.59**). It is inferred that this facies developed in a sheltered, low-energy area.

Carbonate interval 6 is well exposed only along the B 124 road between K38 and K39 (**Figure 3.1**), and is represented by log CA-4b (**Appendix 5**). The coral wacke/floatstone facies is succeeded vertically by coralgal foraminifera float/rudstones and rudstones, culminating with facies association 5 (**Figures 3.44** and **3.60**). This is interpreted as a shallowing upward succession, and represents the increasingly dense development of corals and associated fauna. The presence of a

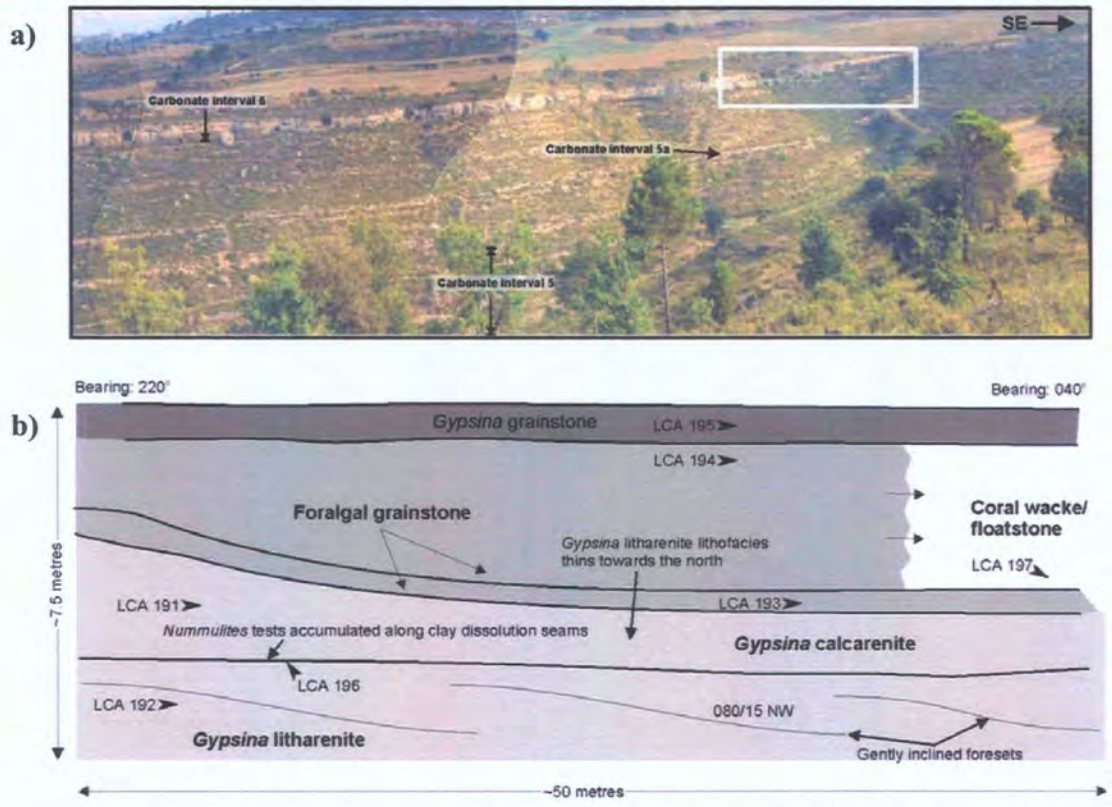


Figure 3.59 Photomontage (a) and close-up field sketch (b) from GR 17352620 (looking NNE) of a clinoform structure at the base of carbonate interval 6. The *Gypsina* calcarenite facies thins towards the northeast over a distance of ~50 m. The overlying foralgal grainstone facies thickens towards the northeast.

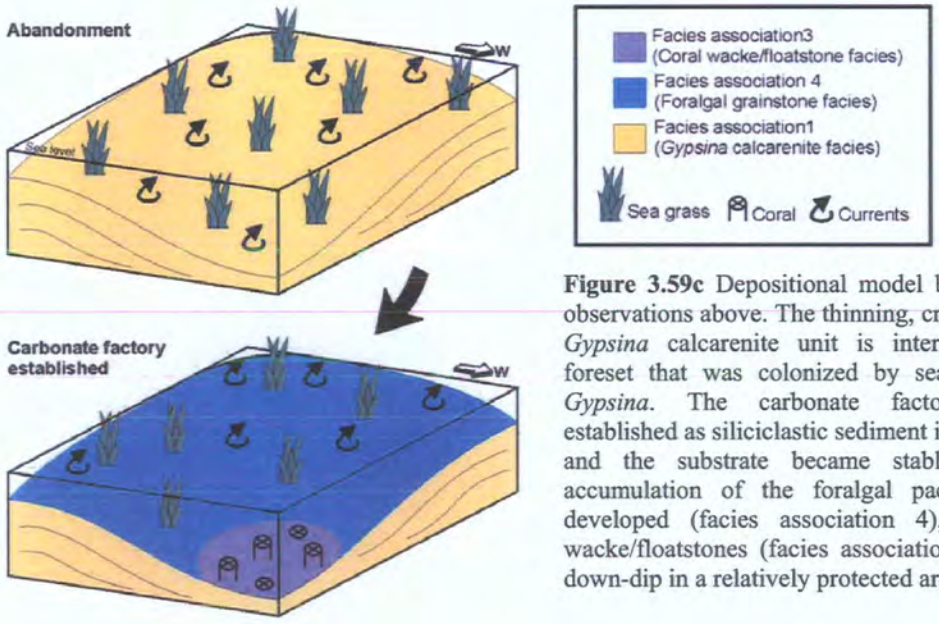


Figure 3.59c Depositional model based on the observations above. The thinning, cross-stratified *Gypsina* calcarenite unit is interpreted as a foreset that was colonized by seagrasses and *Gypsina*. The carbonate factory became established as siliciclastic sediment input reduced and the substrate became stable. A thick accumulation of the foralgal pack/grainstone developed (facies association 4), and coral wacke/floatstones (facies association 3) formed down-dip in a relatively protected area.

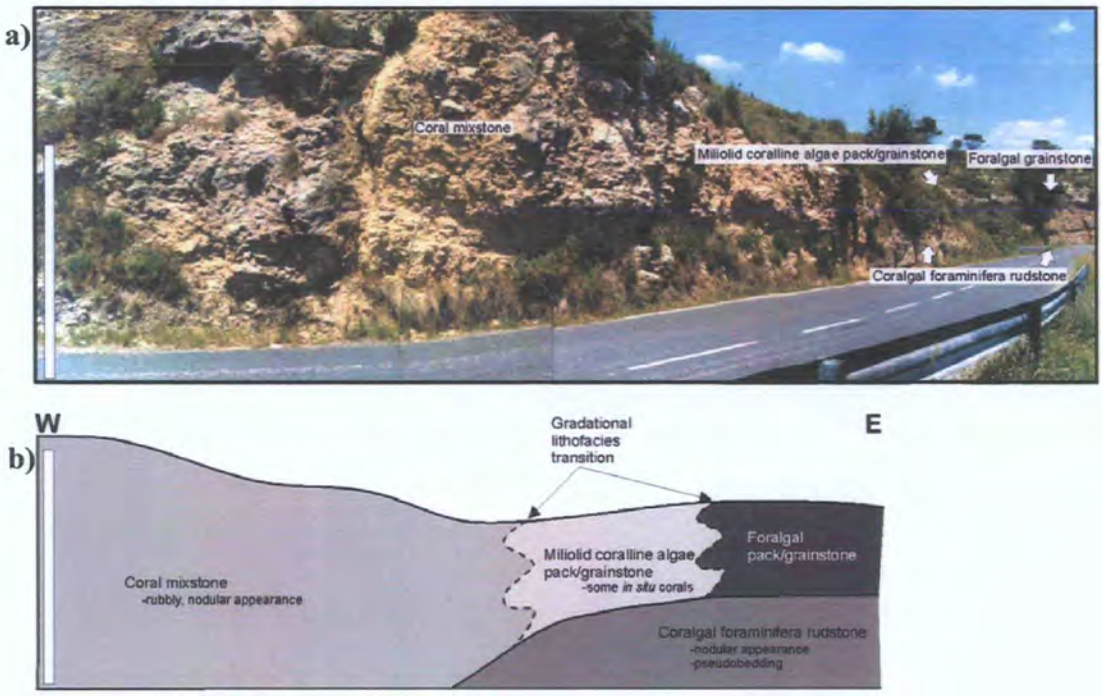


Figure 3.60 Photomontage (a) and interpretation (b) illustrating the relationship between the coral mixstone, coralgal foraminifera rudstone and foralgal grainstones facies (facies associations 5, 3 and 4 respectively) within carbonate interval 6. GR 16142676 (looking NNE). Scale bar = 2 m approximately. Field of view = 50 metres.

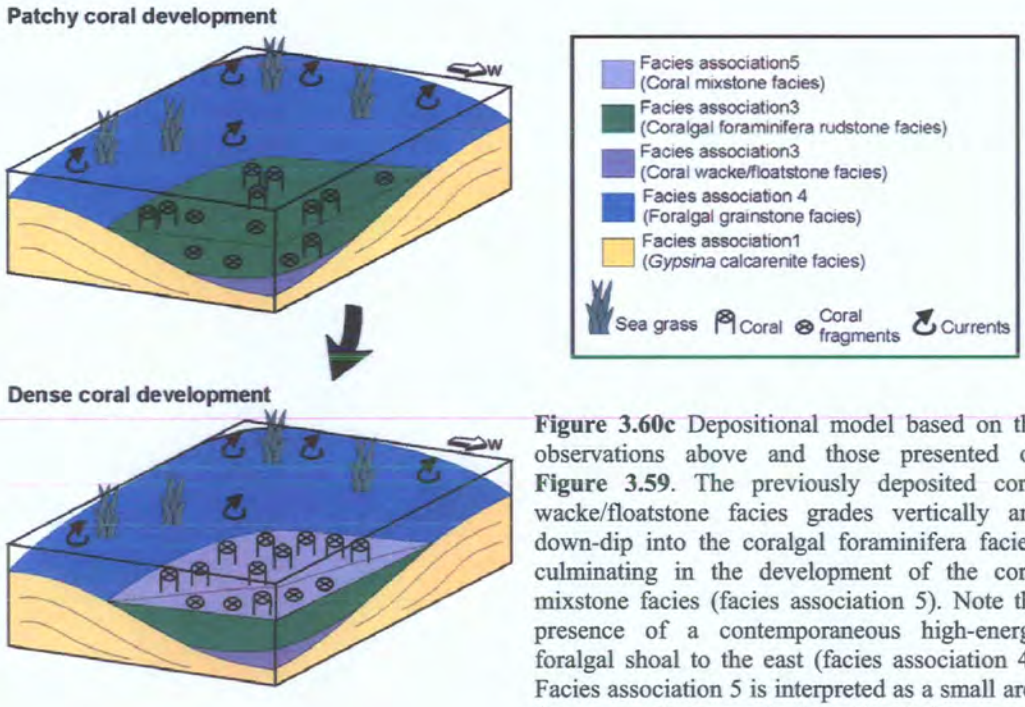


Figure 3.60c Depositional model based on the observations above and those presented on **Figure 3.59**. The previously deposited coral wacke/floatstone facies grades vertically and down-dip into the coralgal foraminifera facies, culminating in the development of the coral mixstone facies (facies association 5). Note the presence of a contemporaneous high-energy foralgal shoal to the east (facies association 4). Facies association 5 is interpreted as a small area (up to ~ 50 metres in length) of dense coral development. See text for discussion.

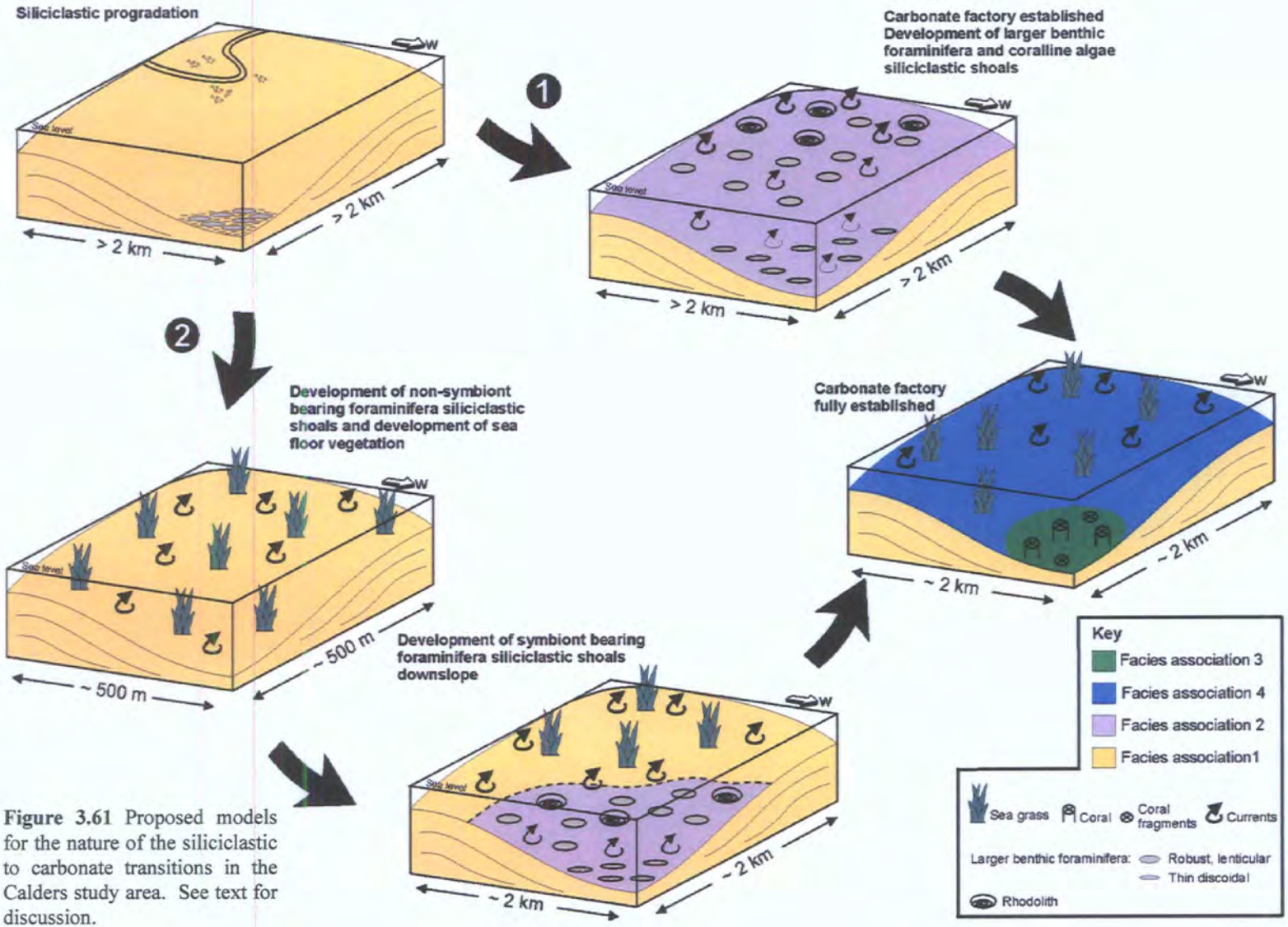


Figure 3.61 Proposed models for the nature of the siliciclastic to carbonate transitions in the Calders study area. See text for discussion.

contemporaneous, high-energy foralgal shoal to the immediate east is inferred from the intercalation of facies association 4 (**Figure 3.60**). The coral mixstone facies is observed to pass laterally into facies association 5 towards the east (**Figure 3.60**).

In summary, the Calders succession represents the deposition of six carbonate-siliciclastic cycles. Sedimentation and the development of biota was influenced by water energy, sediment supply and incident light. Each cycle is represented by an initial siliciclastic lithology with a progradational architecture. Abandonment and stabilisation of siliciclastic macro-bedforms occurred as a consequence of decreased sediment input. The initial biota to develop were tolerant of moderate to high energy levels, sediment input and unstable substrates as marine currents reworked siliciclastics. Biota were dominated by the larger benthic foraminifera *Nummulites, Discocyclina* and *Gypsina,* and coralline algae. With establishment of the carbonate factory and the cessation of coarse siliciclastic input, coral-dominated facies developed. Carbonate development was terminated by a rapid input of siliciclastics that caused the mortality of most biota.

3.5 Sant Amanc facies

The studied units at Sant Amanc are situated along a northeast-southwest trending ridge in the northwestern sector of the study area (**Figure 3.1b**). The units are situated stratigraphically above the carbonate-clastic succession described in **Sections 3.2** to **3.4**. A stratigraphic contact between the carbonates of the main Calders succession and the Sant Amanc succession is not clearly exposed. However, the red mudrocks and sandstones (Artès Formation) and cross-stratified gravelly litharenites and mudrocks (Complejos de Calders Formation) encountered stratigraphically above the Calders section along the N-141 (**Figures 3.3, 3.4** and **3.7**) are also encountered stratigraphically above the units of the Sant Amanc section. Siliciclastic lithologies situated stratigraphically above carbonate interval 6 of the Calders succession are exposed at GR17502641 (log CA-16). A correlation between these sediments and the Sant Amanc succession is presented in **Figure 3.62**.

The exposures at Sant Amanc are orientated northeast-southwest, roughly parallel to the palaeoshoreline trend determined from progradational sedimentary structures in the underlying Calders section. There is no evidence to suggest a

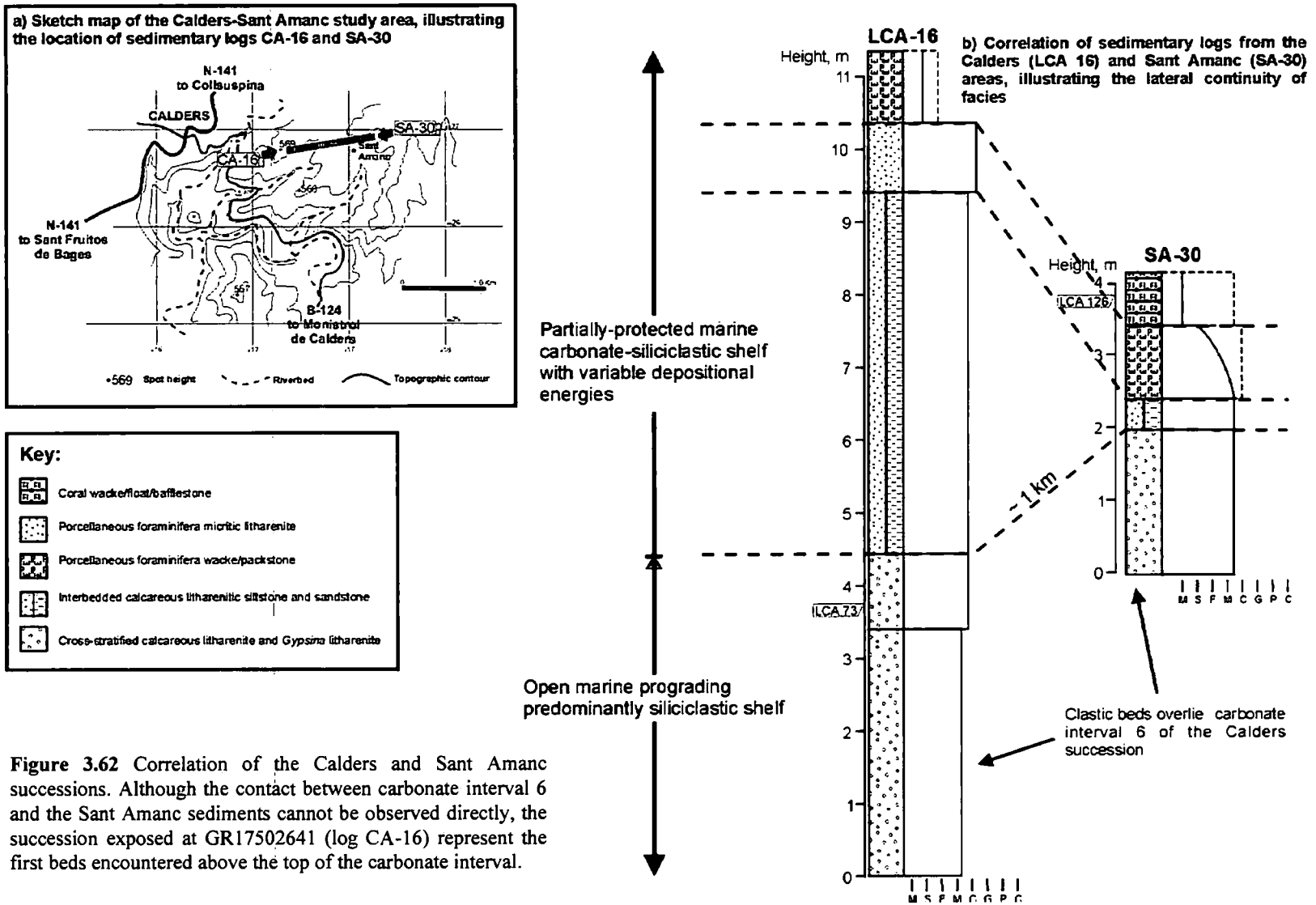
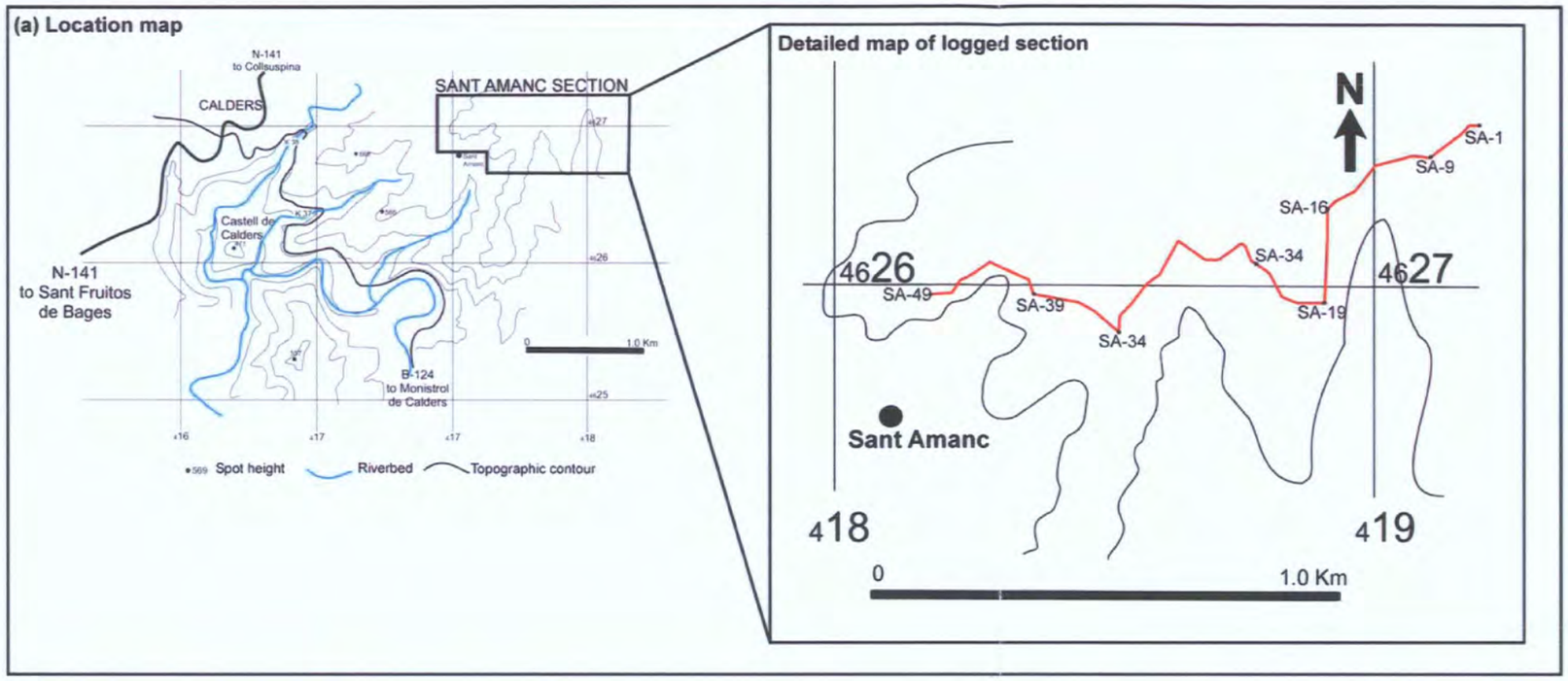
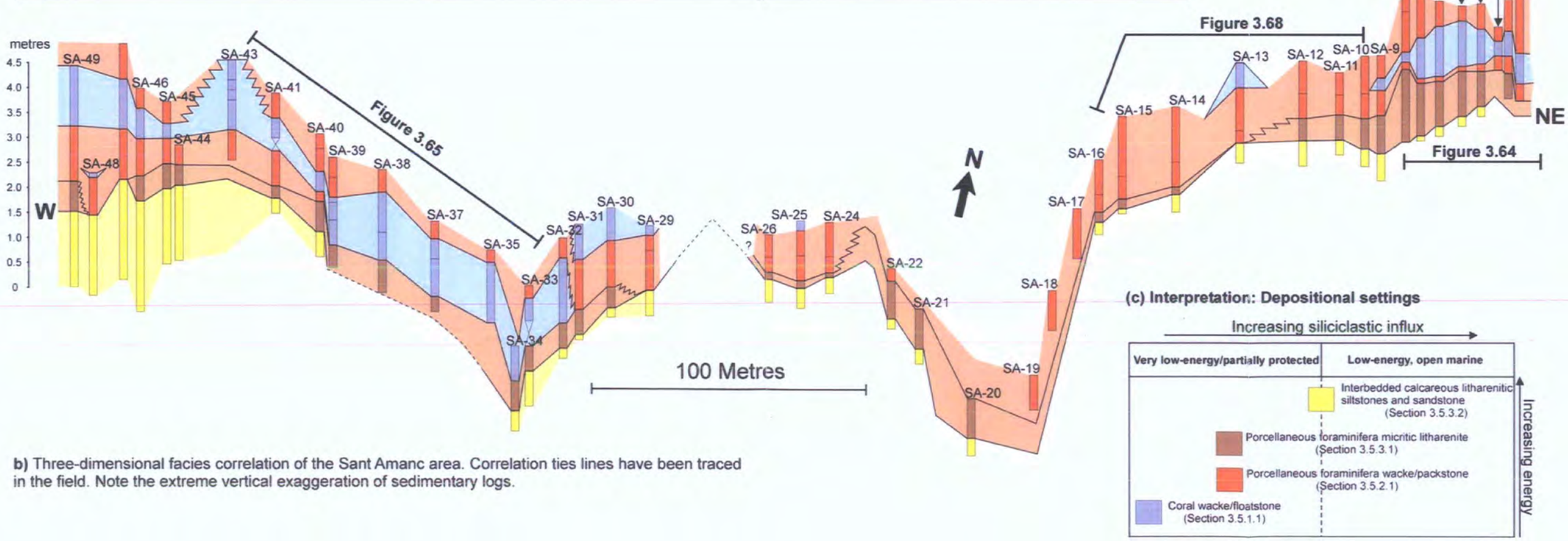


Figure 3.62 Correlation of the Calders and Sant Amanc successions. Although the contact between carbonate interval 6 and the Sant Amanc sediments cannot be observed directly, the succession exposed at GR17502641 (log CA-16) represent the first beds encountered above the top of the carbonate interval.





b) Three-dimensional facies correlation of the Sant Amanc area. Correlation ties lines have been traced in the field. Note the extreme vertical exaggeration of sedimentary logs.

Stratigraphic occurence	Facies	Characteristics	Depositional environment	
Upper Carbonate Unit	Micritic and microsparitic laminated mudstone	Thin beds. MM-scale crinkly laminations of dark micrite and light micrisparite with fenestrae. Low diversity fauna comprising discorbids, small porcellaneous foraminifera and ostracods. Pedogenic nodules.	**Protected, inner-shelf platform**	Bottom colonization by cyanobacterial mats in a restricted intertidal/supratidal setting, possible hypersalinity.
Middle Carbonate Unit	Serpulid packstone	Units thicken over bioherm crests. Fauna dominated by micritic serpulid tubes with rare ostracodes, dasycladaceans, porcellaneous and texturallid foraminifera.		Nutrient-rich, high-energy conditions.
	Porcellaneous foraminifera grainstones alternating with micritic laminated mudstone	Grains dominated by peloids, porcellaneous foraminifera (miliolids, *Spirolina, Taberina*) and texturaliids. Also contains coralline algae (*Corallina, Jania, Lithoporella*), with echinoids, ostracods, bryozoa, bivalves, gastropods, and dasycladacean algae. Crinkly micritic laminae represent cyanobacterial mats.		Shifting bioclast sand shoals periodically fixed by cyanobacterial mats. Environmental conditions alternate between storm and calm intervals.
	Porcellaneous foraminifera and coralline algae grainstones	Megaripples. Variable thickness beds. Foraminifera dominated by *Taberina, Orbitolites, Spirolina* texturaliids and peneroplids. Green (dasyclasdaceans) and red algae (*Lithoporella* and *Corallina* and *Jania*). Oolitic coatings present on some bioclasts.		Current-reworked setting. Shallow conditions with normal marine salinities.
	Crustose coralline algae, bryozoa and attached foraminifera boundstones	Variable relative proportions of articulated coralline algae (*Jania* and *Corallina*), corals, serpulids and siliciclastics. Also encrusting algae (*Lithoporella* and *Lithothamnion*) and foraminifera (victoriellids and haddoniids). Encrusters may form a framework.		Longshore current influenced elongate bioherms. Patchy macro-algae vegetation.
Lower Carbonate Unit	Porcellaneous foraminifera and coralline algae wackestone-packstone	Tabular units, often bioturbated. Highly diverse foraminiferal assemblage dominated by alveolinids, soritids, peneroplids, miliolids, agglutinated forms and valuvulinids. Minor small hyaline forms including discorbids, bolvinids and cibicids. Articulated coralline algae (*Corallina* and *Jania*) and encrusting *Lithoporella*.		Shallow mesotrophic setting with normal marine salinities. Larger porcellaneous foraminifera lived epiphytically on perennial vegetation.

Table 3.4 Summary of facies present in the Eocene "Terminal Complex" sediments of the eastern Pyrenean Foreland Basin, as defined by Travé (1992) and Travé *et al.* (1996).

Table 3.5 Facies scheme for the Sant Amanc area.

Facies group and facies	Dominant textures	Bed thickness	Bedding characteristics	Bed contacts	Lateral continuity	Components	Matrix	Sedimentary structures	Energy level	Environmental interpretation
Coral-dominated facies (**Section 3.5.1**)										
Coral wacke/float/ bafflestone	Wackestone Floatstone Bafflestone Packstone	30 cm to 1.5 m	Lentricular beds, nodular weathering. Nodules up to 15 cm in length. Clay partings.	Sharp to gradational upper and lower.	50 to 100 m	Branching corals, coralline algae, molluscs, larger benthic foraminifera,	Muddy 49.5 to 70 %	Locally aligned nodules interpreted as bedding remnant	Very low	Small, elongate bioherms with long axis elongated normal to the palaeoshoreline. Low nutrients.
Mixed carbonate-siliciclastic facies (**Section 3.5.2**)										
Porcellaneous foraminifera wacke/packstone	Wackestone Packstone	15 cm to 1.5 m	Apparently tabular. Nodular to massive weathering.	Sharp to gradational upper and lower.	> 200 m	*Orbitolites*, *Rhabdorites*, miliolids, coralline algae, molluscs, bryozoa	Muddy 35.5 to 70.5 %	Locally aligned nodules interpreted as bedding remnant	Low to moderate	Vegetated setting, high particulate sedimentation with sporadic siliciclastic input. Locally high nutrient levels
Siliciclastic-dominated facies (**Section 3.5.3**)										
Porcellaneous foraminifera micritic litharenite	Sandstone wackestone	45 cm to 1.45 m	Apparently tabular. Nodular to massive weathering.	Sharp to gradational upper and lower.	> 500 m	*Orbitolites*, *Rhabdorites*, miliolids, coralline algae, molluscs, bryozoa	Muddy 42.5 to 71.5 %	Cross-stratification, bioturbation	Low to moderate	Transitional marginal marine environment. Partially protected with a moderate siliciclastic input.
Interbedd calcareous litharenitic siltstone and sandstone	Sandstone Siltstone wackestone	Up to 5 m	Apparently tabular.	Sharp to gradational upper. Sharp lower.	> 1 km	*Orbitolites*, *Rhabdorites*, miliolids, coralline algae, molluscs, bryozoa. Mainly reworked.	Muddy 51.5 to 64 %	Metre-scale cross stratification, bioturbation	High to low	Shallow marine, above fair weather wave base.

dramatic change in shoreline orientation between deposition of the Calders and the Sant Amanc successions.

Due to its stratigraphic location and the type of facies present, it is interpreted that the Sant Amanc section is a currently unidentified exposure of the "Terminal Complex" defined by Travé (1992). The Terminal Complex is an interval of sediments overlying the Milany Depositional Sequence bounded by two regional uncomformities. The complex represents the final open marine to restricted marine carbonate interval of the eastern Pyrenean Foreland Basin prior to continentalisation and deposition of the Solsona Depositional Sequence (**Chapter 2**). The Terminal Complex is composed of three carbonate units interlayered with prograding siliciclastic sediments. A summary of the characteristics and depositional environments of the terminal Complex is presented in **Table 3.4**.

Lithologies are moderately well exposed within a forested area allowing detailed logging and lateral correlation of beds. Forty-nine sections were logged and correlated (**Figure 3.63**). The facies scheme used to sub-divide the Sant Amanc succession is presented on **Table 3.5**.

3.5.1 Coral-dominated facies

Coral-dominated facies are predominantly carbonate in composition, although they may contain a moderate to low fine-grained siliciclastic component (0 to 10 %). Facies are characterised by *in situ* and reworked colonial corals, and only rarely form a framework. Coral colonies are rarely in contact. Subordinate biota includes porcellaneous foraminifera, coralline algae, molluscs, echinoids and bryozoa. Carbonate textures are floatstone, rudstone, bafflestone and wackestone (**Table 3.4**).

3.5.1.1 Coral wacke/float/bafflestone

Lithologies: Coral wacke/floatstone
Coral float/bafflestone

Occurrence and bed characteristics

This facies is common in the Sant Amanc section. Upper and lower contacts of this facies are sharp to gradational over a few centimetres. Measured bed thickness ranges from 30 cm to 1.5 m. Units are preserved as very low-relief mound like

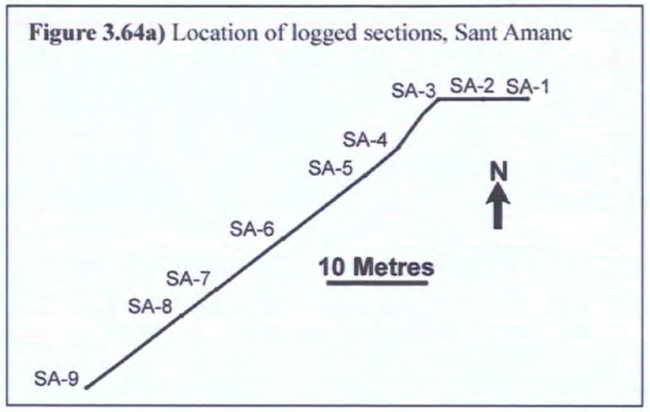
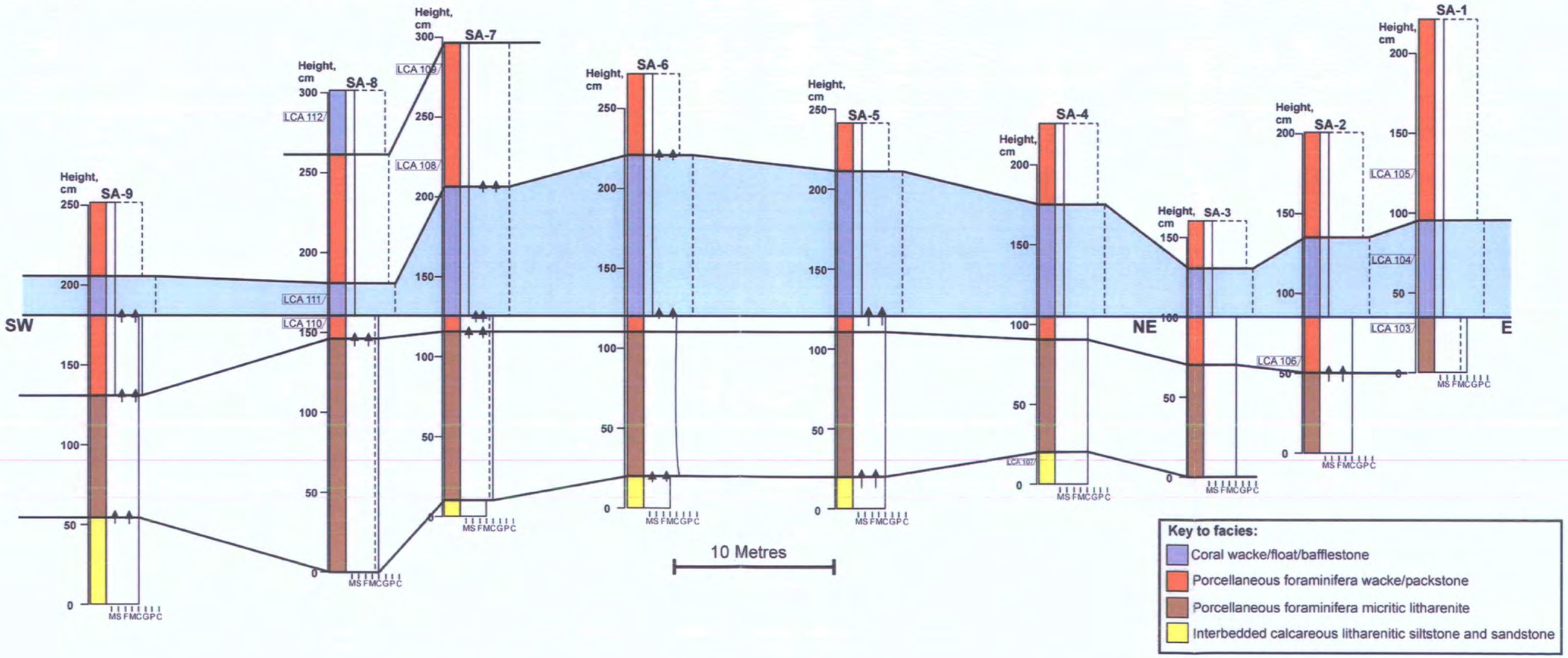


Figure 3.64a) Location of logged sections, Sant Amanc

Figure 3.64b) East-west correlation of logs SA-1 to SA-9 illustrating lateral thickness variations of the coral wacke/float/bafflestone facies. All logs are "hung" from the base of the coral wacke/float/bafflestone facies. Correlation tie-lines have been traced in the field. Note the extreme vertical exaggeration.



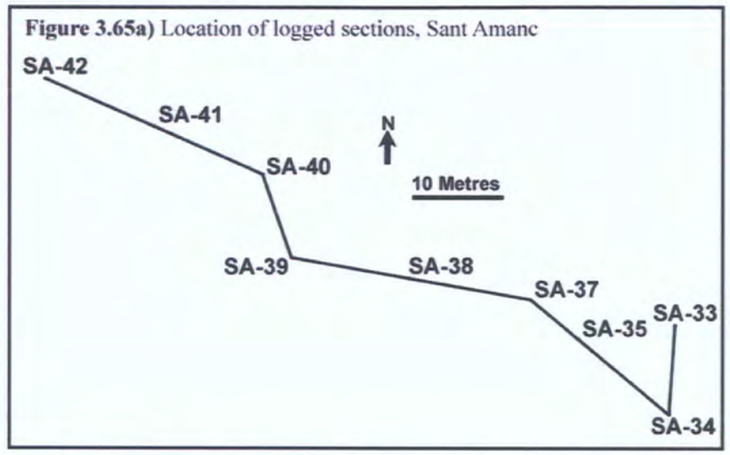
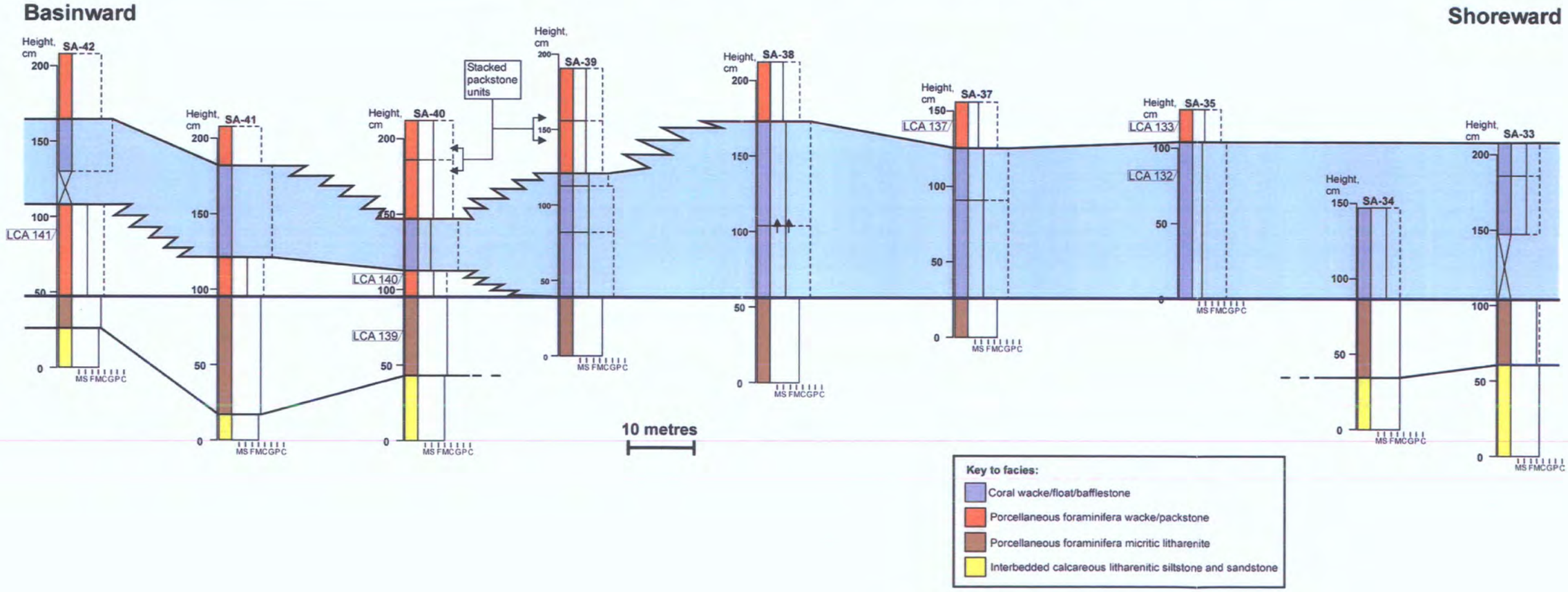


Figure 3.65a) Location of logged sections, Sant Amanc

Figure 3.65b) Northwest-southeast correlation of logs SA-33 to SA-42 illustrating lateral thickness variations of the coral wacke/float/bafflestone facies. All logs are "hung" from the base of the coral wacke/float/bafflestone facies. Correlation tie-lines have been traced in the field. Note the extreme vertical exaggeration.



structures that may be traced laterally for 50 to 100 m (**Figure 3.64**). Two coeval, isolated mounds are identified (**Figure 3.63**). The first is situated towards the northeastern limit of the exposure and is represented by logs SA-1 to SA-9 (**Figure 3.64**). A second larger mound is identified towards the southwestern limit of the exposure, and is represented by logs SA-25 to SA-49. Part of this structure is illustrated on **Figure 3.65**. The coral wacke/float/bafflestone facies often passes laterally into the porcellaneous foraminifera wacke/packstone facies (**Figure 3.65**).

The coral wacke/float/bafflestone facies has a nodular appearance. This facies weathers a pale grey to white colour (**Figure 3.66a**). Fresh surfaces are pale grey to green-blue in colour. Nodules are up to 15 cm and may be locally aligned sub-parallel to bedding. Blue-grey clay-grade material often surrounds nodules, and may form thin (<3 cm) laterally restricted horizons.

Lithological description

This facies is characterised by an abundance of *in situ* and/or fragmented branching and massive corals (**Figure 3.66**). *In situ* corals rarely make up a framework and are only locally in contact with one another. Angular, unabraded coral fragments range from 0.2 mm to 10 cm and comprise 15.5 to 48.5 % of this facies (**Figure 3.66b**). Primary internal cavities are in-filled with matrix and bioclastic debris. Coral fragments occasionally have thin crusts (<1mm) of coralline algae (mainly *Lithoporella* and *Lithothamnion*). Laminar coralline algae crusts comprise up to 5.5 % of this facies. Crusts are often detached and are preserved floating in the matrix.

The coral wacke/float/bafflestone facies contains a moderately diverse faunal assemblage, although fossils typically comprise less than 50 % of this facies. Foraminifera include miliolids (0 to 2 %), *Orbitolites* (0 to 2.5 %), texturaliids (< 1 %) (**Figure 3.67a**), undifferentiated hyaline benthics (a mixture of discorbids, bolvinids and cibicids) (0.5 to 3 %) and encrusting forms such as *Haddonia* and victoriellids (0 to 1.5 %). *Orbitolites* tests, up to 5 mm in diameter, have a flattened, discoidal morphology. Tests show moderate abrasion and fragmentation. Miliolids, dominated by *Quinqueloqulina*, are up to 2 mm in diameter, and are preserved intact although may be abraded. Red staining of porcellaneous foraminifera tests is observed in some samples of this facies.

Other fossils present in this facies include gastropods (0.5 to 3 %), bivalves (0.5 to 2 %), echinoid spines and plates (0.5 to 3.5 %), peloids (possibly highly abraded coralline algae fragments) (up to 11.5 %) and very rare bryozoa (< 0.5 %). Delicate bivalves up to 2.5 cm are preserved disarticulated and unfragmented. Ornamented gastropods (*Cerithium sp.*) up to 2 cm are typically preserved intact **(Figure 3.67b)**.

Dark-brown micritic matrix comprises 49.5 to 70 % of this facies **(Figures 3.66b and 3.67b)**. The matrix is interpreted to contain some clay-grade organic and/or terrigenous material by its dark, locally patchy nature. Silt-grade quartz comprises 0 to 2.5 % of this facies. Quartz grains are well sorted, angular and subspherical.

Diagenesis

Cementation of this facies is inferred by its nodular appearance, although it is difficult to resolve carbonate cement from the fine, muddy matrix. Formerly aragonitic fossils (molluscs and corals) have been leached with the resultant biomouldic pores fully occluded with drusy calcite spar cement **(Figure 3.67b)**. Sutured nodule contacts are observed in some localities, and blue-grey circum-nodular clays indicate dissolution and concentration of insolubles. Milimetre-scale drusy calcite spar cemented microfractures are observed in thin section **(Figure 3.67a)**. Blue-grey mud surrounding nodules is interpreted as circum-nodular dissolution seams and their formation is likely related to post-depositional compaction and dissolution of the succession.

Interpretation: depositional environment

The coral wacke/float/bafflestone facies contains a marine faunal assemblage. Normal marine salinities are inferred from the presence of corals and *Orbitolites* (Ghose 1977, Leutenegger 1984, Geel 2000). Recent larger porcellaneous foraminifera comparable to *Orbitolites* may inhabit high-salinity settings, although they tend to adopt aberrant morphologies that are not observed in this facies (Reiss and Hottinger 1984). *Orbitolites* is an endosymbiont-bearing foraminifera, thus deposition within the photic zone is interpreted from its presence (Leutenegger 1984, Geel 2000).

a)

b)

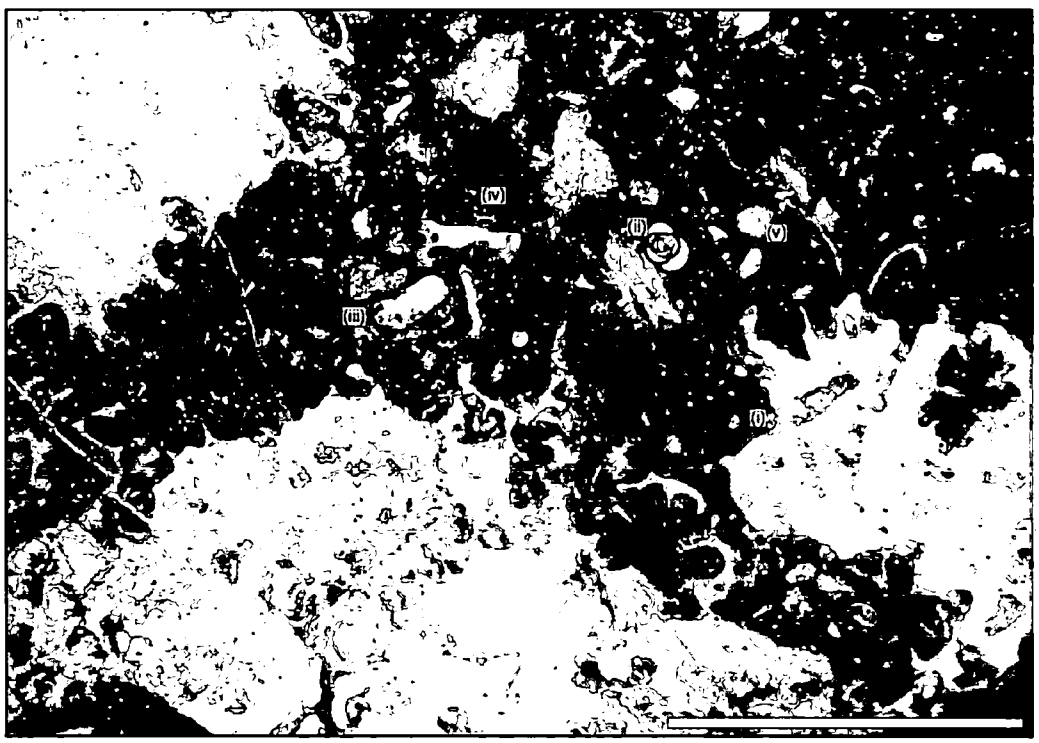


Figure 3.66a Field exposure of the coral wacke/float/bafflestone facies, log SA-32, bed 3. Hammer for scale=45cm.

Figure 3.66b Photomicrograph (PPL) of the coral wacke/float/bafflestone facies (Sample LCA 110, log SA-8, bed 3). (i) Large coral fragments, (ii) miliolid, (iii) agglutinated foraminifera, (iv) rounded coralline algae fragment, (v) echinoid spine. Scale bar=2mm.

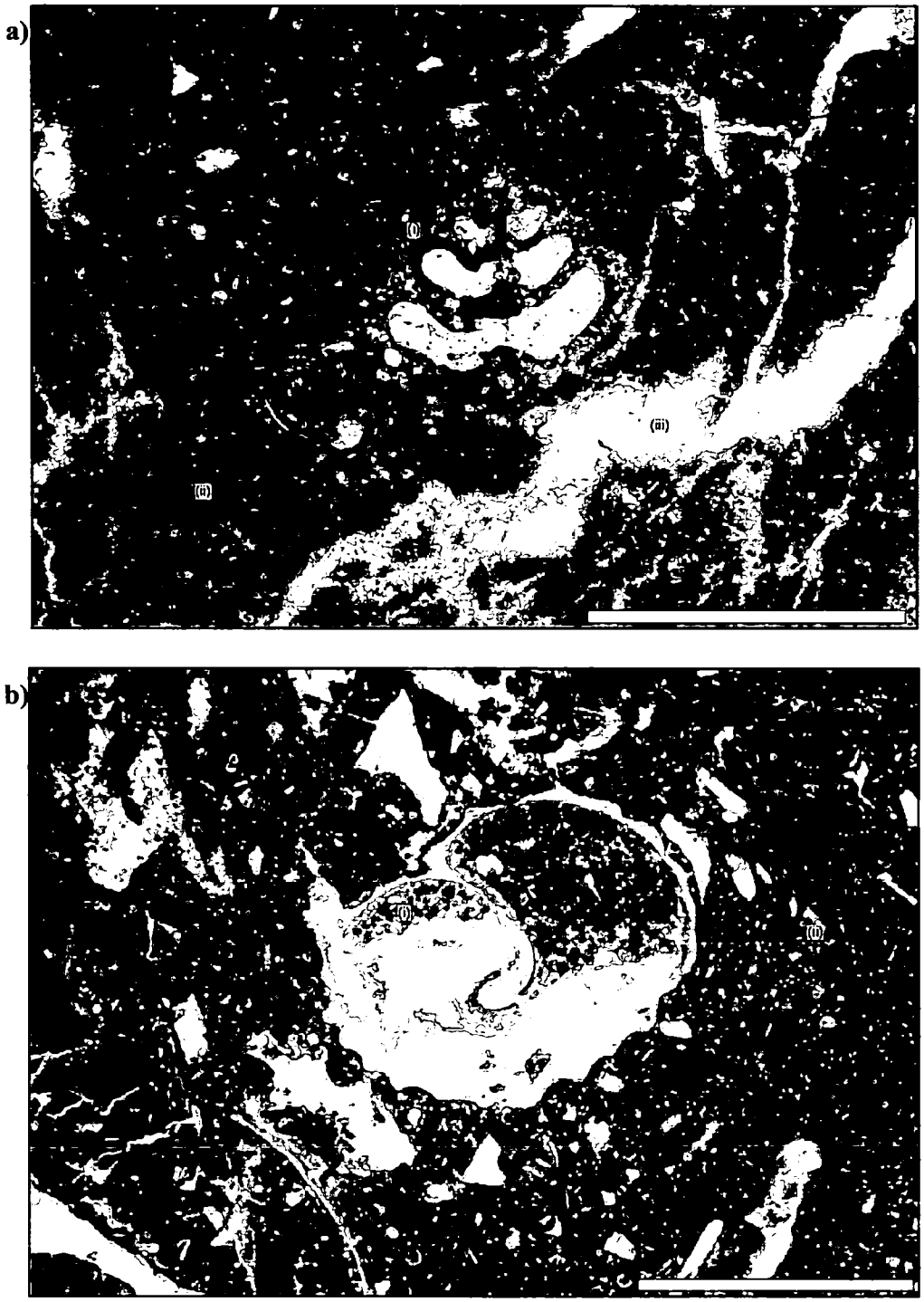


Figure 3.67a Photomicrograph (PPL) of the coral wacke/float/bindstone lithofacies. Sample LCA 145, log SA-44, bed 4). (i) texturaliid foraminifera, (ii) dark brown-green muddy matrix which comstitutes upto 70% of this lithofacies, (iii) fine dissolution seams in-filled with calcite. Scalebar=1mm.

Figure 3.67b Photomicrograph (XPL) of the coral wacke/float/bafflestone lithofacies. Sample LCA 147, log SA-47, bed 3). (i) Intact, ornamented *Cerithium* gastropod. Note that the shell is composed of secondary calcite spar and retains no primary structures, (ii) fine skeletal hash present within the dark muddy matrix. Scalebar=2mm.

Porcellaneous foraminifera, such as miliolids and *Orbitolites,* often adapt an epiphytic lifestyle, living on sea grasses and/or macro-algae (Chaproniere 1975, Hallock and Peebles 1993, Geel 2000). The presence of perennial sea floor vegetation is inferred from the occurrence of large discoidal *Orbitolites* that are thought to have had a long (annual) lifespan. The baffle capacity of this vegetation may have promoted the accumulation of carbonate mud. Larger encrusting foraminifera (*Haddonia* and victoriellids) are typical of cryptic environments within the coral debris or on vegetation rhizomes feeding on organisms present in interstitial waters (Reiss and Hottinger 1984). In this facies, however, they are largely detached. The dearth of large encrusting foraminifera suggests that their environmental requirements were not met, and colonisation may have been inhibited by the abundance of intergranular mud (burying benthic habitats).

Low-nutrient conditions are concluded from the presence of corals and symbiont-bearing larger foraminifera. Local nutrient-rich areas must have existed however to support the moderate array of deposit feeders and grazers (i.e. bivalves, gastropods, echinoderms etc) and non-symbiont bearing small benthic foraminifera. It is suggested by Kitazato (1988) that a moderate abundance of small, free-living benthic foraminifera indicates relatively eutrophic conditions at the sediment-water interface where nutrients are recycled through burrowing.

Low-energy conditions are concluded from the high percentage of fine-grained matrix and the well-preserved nature of delicate bivalves and gastropods. Thin, laterally extensive coralline algae crusts are also indicative of lower-energy marine environments where competition for space is limited (Reid and MacIntyre 1988, Minnery *et al.* 1985, Minnery 1990). A terrigenous influence is inferred from the presence of silt-grade quartz. It is inferred from their angular nature that siliciclastic grains were deposited from suspension and underwent limited post-depositional reworking and/or abrasion prior to deposition and lithification. It is suggested that fragmented and abraded bioclasts such as coralline algae are derived from a relatively high-energy environment, as other bioclasts in the coral wackestone floatstone facies are relatively well preserved.

The coral wacke/float/bafflestone facies occurs as very low-relief patch reef accumulations, interpreted as biostromes. Log correlations demonstrate that the long axis (and maximum thickness) of coral wacke/float/bafflestone units is orientated

roughly perpendicular to the palaeoshoreline (**Figure 3.63**). It is inferred from this that offshore-directed currents influenced the final morphology of units of this facies. This is in contrast to Travé (1992) and Travé *et al.* (1996) who demonstrate that longshore currents strongly influenced the morphology of coralgal bioherms of the Terminal Complex exposed in the eastern Vic Basin.

In summary, it is interpreted that the coral wacke/float/bafflestone facies represents coral colonisation in a low-energy, mud-rich environment with open circulation, probably close to or below fair-weather wave base. Beds of this facies are interpreted as very low relief bioherms. Thickness variations across individual facies units imply some topographic relief, although there is no evidence to suggest that true, framework-dominated patch reefs with significant relief existed. This facies has affinities with SMF 8 of Wilson (1975) and the Coral Wackestone facies of Sayer (1995).

3.5.2 Mixed carbonate-siliciclastic facies

Facies included in this grouping are predominantly carbonate in composition, although fine to coarse sand-grade siliciclastic material may constitute up to 50 %. Facies are characterised by the presence of larger porcellaneous benthic foraminifera, in particular *Orbitolites, Rhabdorites* and *Quinqueloculina.* Subordinate biota includes coralline algae, smaller rotaline benthic foraminifera, molluscs, echinoids and bryozoa. The dominant textures observed are siliciclastic packstones and wackestones (**Table 3.5**).

3.5.2.1 Porcellaneous foraminifera wacke/packstone

Lithologies: Porcellaneous foraminifera siliciclastic packstone
Porcellaneous foraminifera wacke/packstone
Miliolid wacke/packstone
Bioclastic packstone

Occurrence and bed characteristics

This facies is very common in the Sant Amanc study area. Units of the porcellaneous foraminifera wacke/packstone facies can be traced laterally for at least 200 m along the length of the ridge (**Figure 3.63**). Measured bed thickness ranges

from 15 cm to 1.25 m. This facies is also identified in an isolated exposure situated to the west of the Sant Amanc ridge (**Figure 3.62**).

Surfaces of the porcellaneous foraminifera wacke/packstone facies weather to a grey-brown colour. Fresh surfaces are a pale-grey to green colour. Beds of this facies are nodular to massive. Nodules are up 15 cm in length and may be surrounded by a soft grey-green siltstone/mudstone. Nodules are commonly aligned forming pseudobedding. Pseudobeds are parallel to bedding contacts. Vertical and sub-horizontal *Thalassanoides*-type burrows up to 5 cm in length are observed in some beds (e.g. log SA-7, bed 3).

Upper and lower bedding contacts may be sharp to diffuse. Lower bedding contacts with the porcellaneous foraminifera micritic litharenite are gradational, and the relative percentage of siliciclastic material decreases vertically from up to 21 % at the contact (e.g. LCA 149, log SA-49) to less than 1 % a few centimetres from the contact (e.g. LCA 106, log SA-2). Upper and lower contacts of the porcellaneous foraminifera wacke/packstone facies with the coral wacke/float/bafflestone facies are often gradational (e.g. log SA-9, beds 3 and 4; log SA-39, beds 3 and 4) but may be sharp (e.g. log SA-32, beds 3 and 4).

The porcellaneous foraminifera wacke/packstone facies is also observed to grade laterally into the porcellaneous foraminifera micritic litharenite facies (e.g. logs SA-22 and SA-24) and also the coral wacke/float/bafflestone facies (e.g. logs SA-8 and SA-9) over distances of around 10 metres (**Figure 3.68**).

Lithological description

This facies is characterised by a moderate abundance and diversity of porcellaneous benthic foraminifera including *Orbitolites* (0 to 5.5 %), *Rhabdorites* (0 to 5 %), miliolids (1.5 to 10 %) and smaller hyaline benthics (0 to 5 %) (**Figures 3.69a** and **b**). Foraminifera contribute 6 to 19.5 % of samples of this facies. *Orbitolites* tests range from 2 mm to 1 cm. Tests have a flattened, discoidal morphology and are typically preserved intact and unabraded (**Figure 3.69a**). *Rhabdorites* tests have an extended, conical morphology. Tests are up to 6 mm, are typically preserved intact although they may be abraded (**Figure 3.69a** and **3.70a**). Miliolid tests are up to 5 mm and are typically preserved intact, although some tests

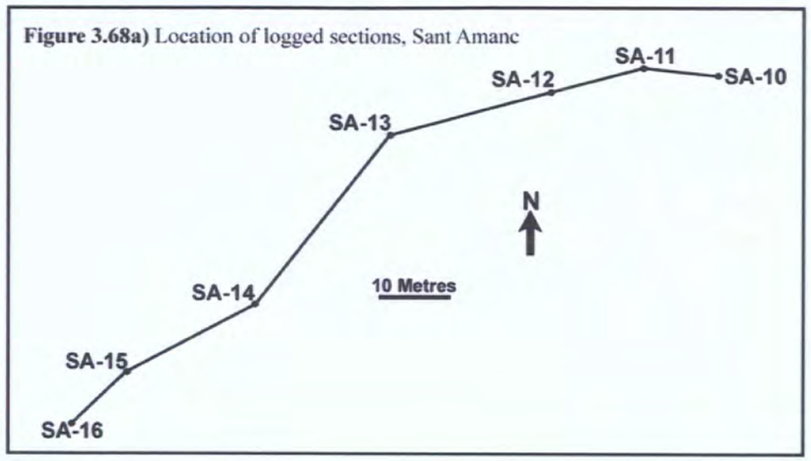
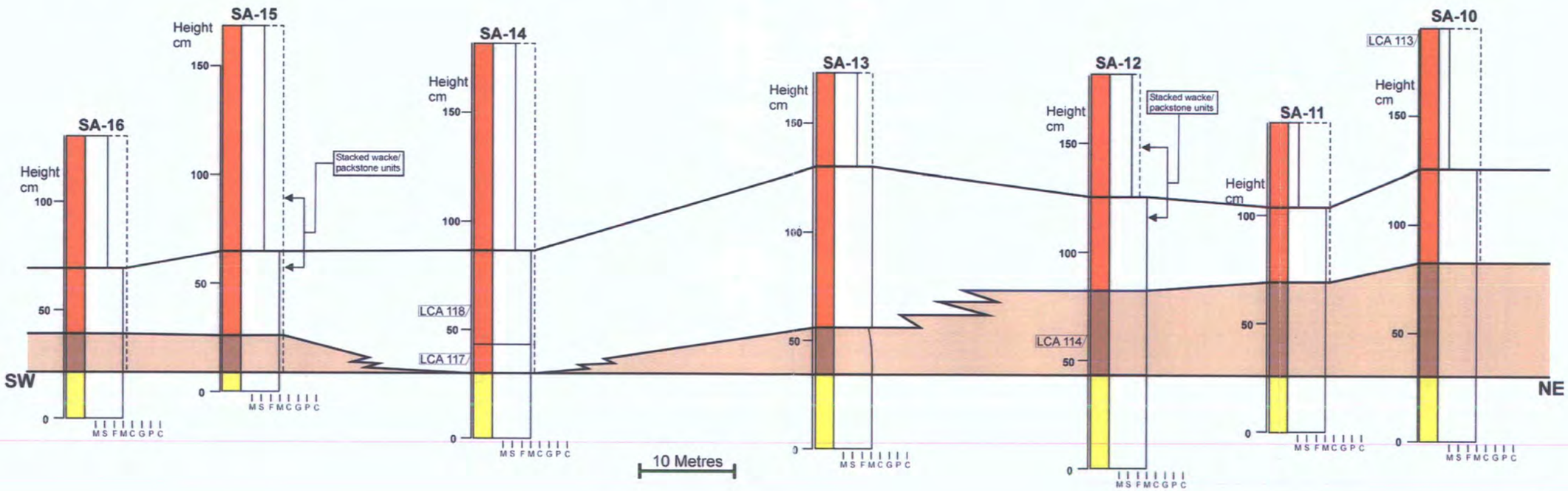


Figure 3.68a) Location of logged sections, Sant Amanc

Figure 3.68b) Northeast-southwest correlation of logs SA-10 to SA-16 illustrating the relationship between the porcellaneous foraminifera wacke/packstone and porcellaneous foraminifera micritic litharenite facies. All correlation tie-lines have been traced in the field. Note the vertical exaggeration.



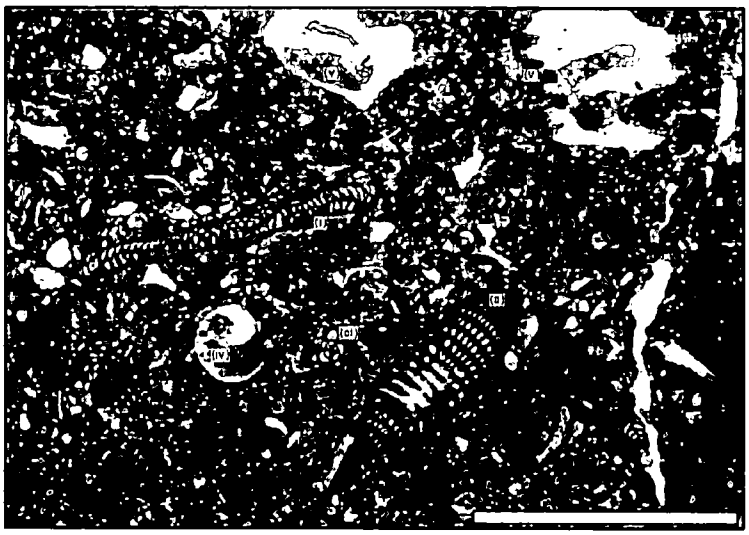


Figure 3.69a Photomicrographs (all PPL) of the porcellaneous foraminifera wacke/packstone lithofacies. a) Sample LCA 105, log SA-1, bed 3. (i) Flattened *Orbitolites* test, (ii) *Rhabdorites,* (iii) *Quinqueloculina,* (iv) intact *Cerithium* gastropod. Note the abundance of fine-grained, muddy matrix. Scalebar =2mm.

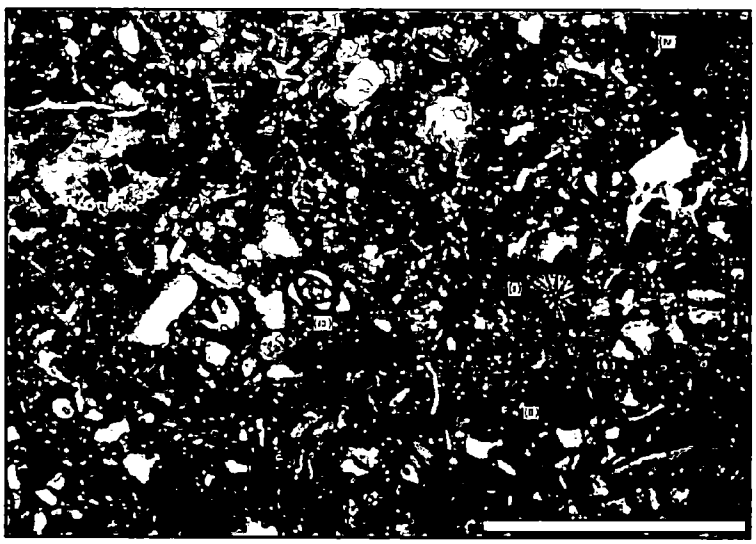


Figure 3.69b. Sample LCA 133, log SA-35, bed 2 (i) *Rhabdorites,* (ii) rounded coralline algae fragment, (iii) slightly reddened miliolid test. Scalebar=2mm.

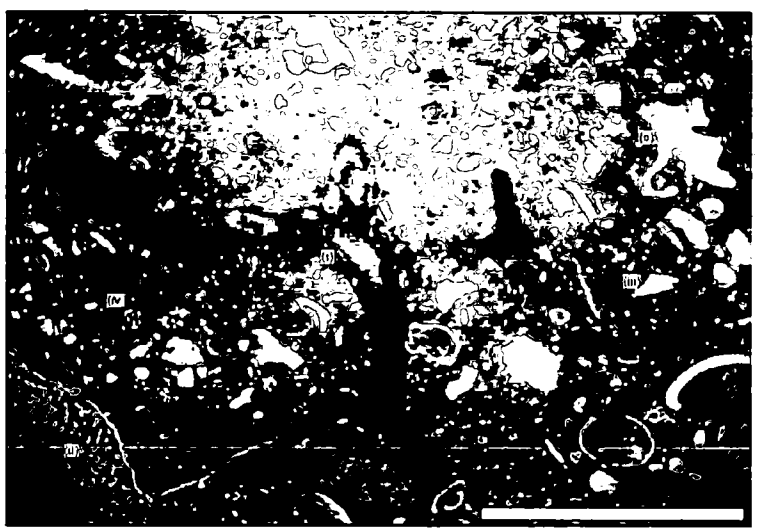


Figure 3.69c. Sample LCA 109, log CA-7, bed 5 (i) branching coralline algae (ii) coral fragments, (iii) angular siliciclastic grain (iv) miliolid. Scalebar=2mm.

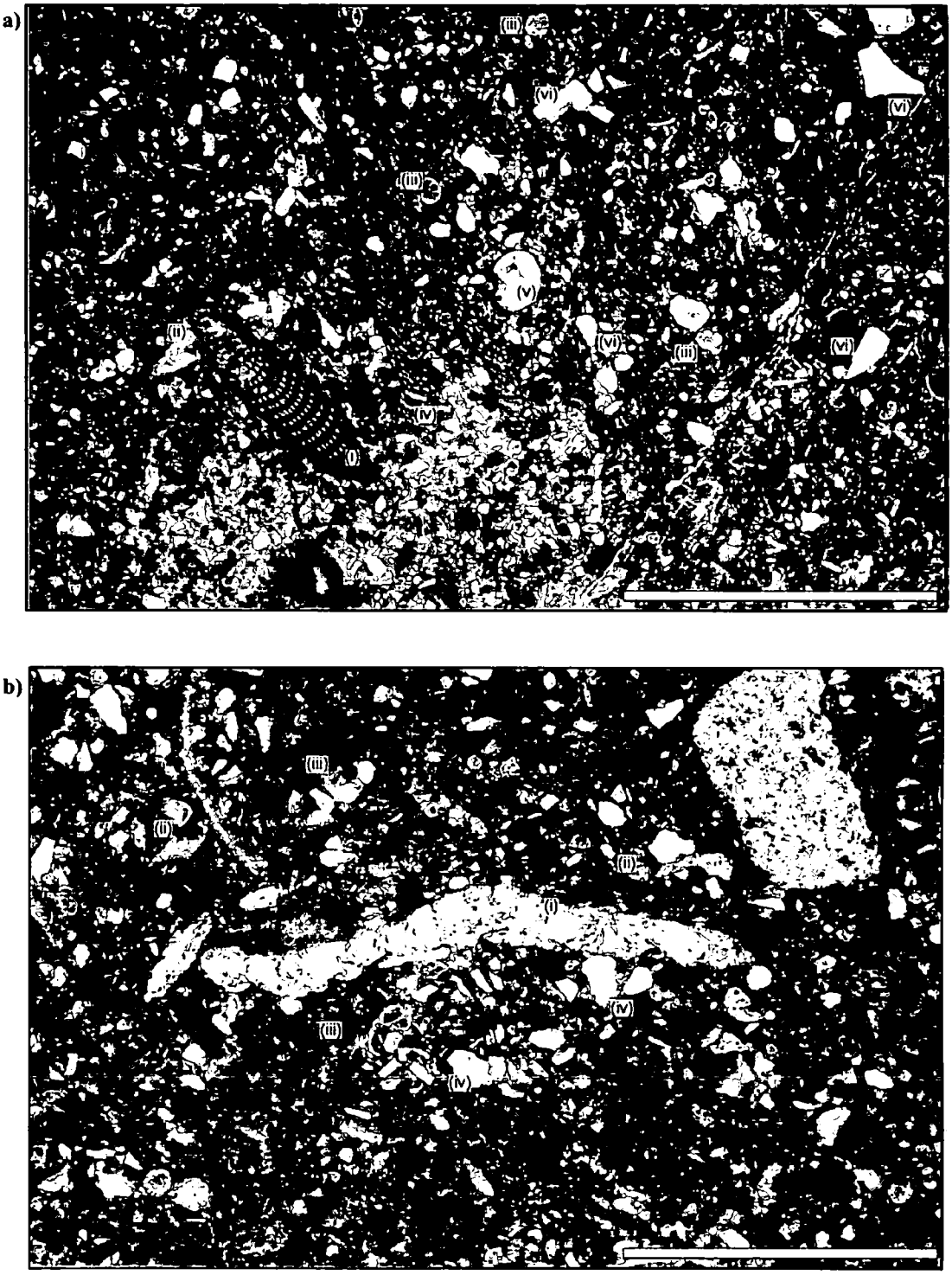


Figure 3.70 Photomicrographs (PPL) of relatively siliciclastic-rich samples of the porcellaneous foraminifera wacke/packstone lithofacies. **a)** Sample LCA 117, log SA-14, bed 3 (i) *Rhabdorites*, (ii) bolvinid small benthic foram, (iii) discorbids, (iv) *Orbitolites*, (v) ostracod, (vi) angular clastic grains. **b)** Sample LCA 123, log SA-30, bed 2 (i) *Planolinderina*, (ii) discorbids, (iii) peloid, probably highly-abraded coralline algae, (iv) clastic grains. Scalebar=2mm.

have a slightly reddened appearance. Other foraminifera present in lesser numbers are *Spirolina, Valvulina, Planolinderina* (**Figure 3.70b**) and a moderate diversity of small hyaline benthics (discorbids and rare bolvinids). Discorbids are particularly abundant in relatively clastic-rich samples (e.g. LCA 117, log SA-14) (**Figure 3.70a**).

Other fossils present in this facies include delicate fragmented branching corals (0 to 7.5 %), gastropods (0 to 2%), bivalves (0 to 8 %), fenestrate bryozoa (< 0.5 %), echinoid spines and plates (0 to 5.5 %), coralline algae (0 to 6.5 %), ostracods (0 to 0.5 %), brachiopods (< 0.5 %) and undifferentiated skeletal debris (1.5 to 12.5 %) (**Figures 3.69a** and **b**). Coralline algae occur as fragmented, well-abraded laminar (crustose) and articulated forms up to 1.5 cm in length, and peloids (≤ 15 %) (**Figure 3.69c**). Primary structures are poorly preserved and fragments are abraded. Coral fragments, up to 4 cm in length, are always fragmented and abraded (**Figure 3.69c**). Bivalves and ornamented gastropods are typically preserved intact and have thin walled shells.

Dark-brown micritic matrix comprises 35.5 to 75.5 % of this facies. Evidence for bioturbation in thin section is provided through localised concentrations of bioclasts and diffusely peloidal matrix (e.g. sample LCA 140, log SA-40). Siliciclastic material comprises < 0.5 to 21 % of samples of this facies. Siliciclastic grains identified are predominantly detrital quartz (mono- and polycrystalline varieties) with a lesser abundance of lithics (quartz-mica schist and siltstones) and calcite (**Figure 3.69c**). Grainsize ranges from 0.1 to 0.5 mm. Siliciclastics are moderately well sorted, angular to subangular and subspherical.

Diagenesis

Originally aragonitic bioclasts (corals, bivalves and gastropods) have been leached, and the resultant mouldic macropores have been completely occluded with calcite spar cement. Foraminifera chambers have been cemented with poikilotopic calcite. Recrystallisation of the micrite matrix is evident in some samples (e.g. LCA 123, log SA-9), although this is atypical of this facies. Extensive cementation of this facies is inferred by its nodular appearance in the field, although it is difficult to resolve carbonate cement from the fine, muddy matrix. Sutured nodule contacts are observed in some localities, and mm-scale microsparitic seams are observed in thin section.

Interpretation: depositional environment

The porcellaneous foraminifera wacke/packstone facies contains a marine faunal assemblage. Normal marine salinities are inferred from the presence of large discoidal *Orbitolites* rather than the abberant form more typical of higher salinity settings (Leutenegger 1984, Geel 2000). Depostion within the photic zone is inferred from the presence and well-preserved nature of the symbiont-bearing foraminifera *Orbitolites, Rhabdorites* and *Spirolina* (Leutenegger 1984, Geel 2000).

The presence of perennial sea floor vegetation (i.e. sea grasses, macroalgae etc.) is inferred from the relative abundance of the large epiphytic foraminifera *Orbitolites* and miliolids, although miliolids are also common in benthic habitats (Reiss and Hottinger 1984, Hallock and Peebles 1993). Recent miliolids such as *Quinqueloculina* and *Triloculina* feed on the superficial algal-bacterial film on vegetation (Kitazato 1988). Articulated and branching coralline algae are moderately abundant in this facies and are thought to have been a suitable substrate. It is postulated that sea grasses were present, but as they contained no hard parts, their preservation in the rock record would be unlikely. The baffle capacity of this vegetation may have promoted the accumulation of carbonate mud that is the most significant component of this facies (cf. Porcellaneous foraminifera and Coralline Algae Wackestone-packstone facies of Travé (1992) and Travé *et al.* (1996)).

Oligotrophic conditions are deduced from the presence of large, symbiont-bearing foraminifera (Reiss and Hottinger 1984). However, the possibility of localised nutrient-rich areas is suggested by the abundance of deposit feeding and burrowing organisms, and non-endosymbiont bearing smaller benthic foraminifera (discorbids and bolvinids). Travé (1992) and Travé *et al.* (1996) attribute a mixture of larger foraminifera with burrowing organisms to the presence of a thin eutrophic layer at the sediment-water interface. The majority of the thickness of the water column is however, oligotrophic.

There is no evidence of *in situ* coral growth within the porcellaneous foraminifera wacke/packstone facies although large coral fragments and encrusting foraminifera (*Planolinderina*) were derived locally. The development of significant coral accumulations may have been inhibited in this environment by the moderately high nutrient levels (discussed above) as corals favour nutrient-poor conditions.

Proximity to a siliciclastic source (i.e. the coastline) is inferred from the moderate abundance of quartz and lithic grains in this facies. Limited abrasion due to transport and re-working prior to deposition and lithification is interpreted from the angular nature of grains. Low-energy conditions and an absence of winnowing currents are inferred from the abundant fine-grained matrix.

In summary, the porcellaneous foraminifera wacke/packstone facies is interpreted to represent a low-energy, partially protected (but not isolated) marine environment, close to the coastline. Siliciclastic sedimentation rates were low allowing substrate colonisation by calcareous benthic organisms. Available substrates were also provided on vegetation. This facies has affinities with the Porcellaneous Foraminifer and Coralline Algae Wackestone Packstone facies of Travé (1992) and Travé *et al.* (1996). This facies is interpreted as a mesotrophic, partially protected inner platform setting. In comparison, there was a higher siliciclastic input in the porcellaneous foraminifera wacke/packstone facies of the Sant Amanc section.

3.5.3 Siliciclastic facies group

Facies included in this grouping are predominantly siliciclastic in composition, although they may contain a significant carbonate component as a combination of bioclasts, cement and detrital carbonate grains. The dominant textures are sandstone and pebbly sandstone. Grains are well-sorted, sub-angular and sub-spherical. Biota may be dominated by benthic foraminifera including *Orbitolites, Rhabdorites* and *Quinqueloculina*, with molluscs, echinoids and serpulids (**Table 3.5**).

3.5.3.1 Porcellanaeous foraminifera micritic litharenite

Lithologies: Porcellaneous foraminifera micritic litharenite
Porcellaneous foraminifera litharenite

Occurrence and bed characteristics

Stratigraphically, the porcellaneous foraminifera micritic litharenite facies always occurs towards the base of the exposed succession, above the interbedded calcareous siltstone and sandstone and below the porcellaneous foraminifera wacke/packstone and coral wacke/float/bafflestone facies (**Figure 3.63**). The lateral extent of the porcellaneous foraminifera micritic litharenite facies cannot be

determined in the field due to the lack of continuous field exposure. However, this facies can be traced across the extent of the Sant Amanc area and thus its lateral extent exceeds 500 m. This facies was also encountered in an isolated exposure situated to the west of the main ridge (log CA-16, GR 17502641), suggesting a lateral extent exceeding 1 km.

Measured bed thickness of this facies ranges from 45 cm to 1.45 m. Upper and lower bedding contacts are gradational (e.g. log SA-6, beds 1 to 3). The porcellaneous micritic litharenite facies is observed to grade laterally into the porcellaneous foraminifera wacke/packstone facies (e.g. logs SA-12 and SA-12) over distances of less than 20 m.

Fresh and weathered surfaces are a pale-brown to buff colour (**Figure 3.71a**). Beds may be nodular, with nodules ranging from 10 to 15 cm in length. Nodules are often aligned forming pseudobedding (**Figure 3.62a**). Pseudo-crossbeds dip consistently northward (logs SA-7 bed 2, SA-9 bed 2), normal to the inferred palaeoshoreline orientation and consistent with palaeotransport directions measured from the underlying Calders succession. Beds may also be characterised by the presence of vertical and horizontal *Skolithos* and *Thalassanoides*-type burrows (e.g. log SA-20 beds 1 and 2, log SA-32 bed 2). Burrows are up to 10 cm in length.

Lithological description

This facies is a matrix-supported sandstone with variable bioclast content. Siliciclastic grains are angular, moderately well to poorly sorted and subspherical. Grainsize ranges from 0.1 to 0.5 mm. Siliciclastic grains identified are mono- and polycrystalline quartz (10 to 35.5 %), lithics (2 to 15.5 %), and feldspar (0 to 1 %).

Lithic grains are siltstones and quartz-mica schist. Micas may have a 'rotted', appearance, and are altered to a yellow to rusty-brown coloured clay observed in thin section. Grains are supported within a dark brown, micrite matrix, comprising 42 to 71.5 % of this facies (**Figures 3.71b** and **c**).

Bioclasts are angular and poorly sorted, comprising up to 34.5 % of this facies. This facies contains a diverse benthic foraminifera assemblage with foraminifera comprising up to 18.5 % of a sample. Foraminifera present include *Orbitolites* (0 to 3.5 %), *Rhabdorites* (≤0.5 %), miliolids (1 to 10 %), *Spirolina* (0 to 1.5 %),

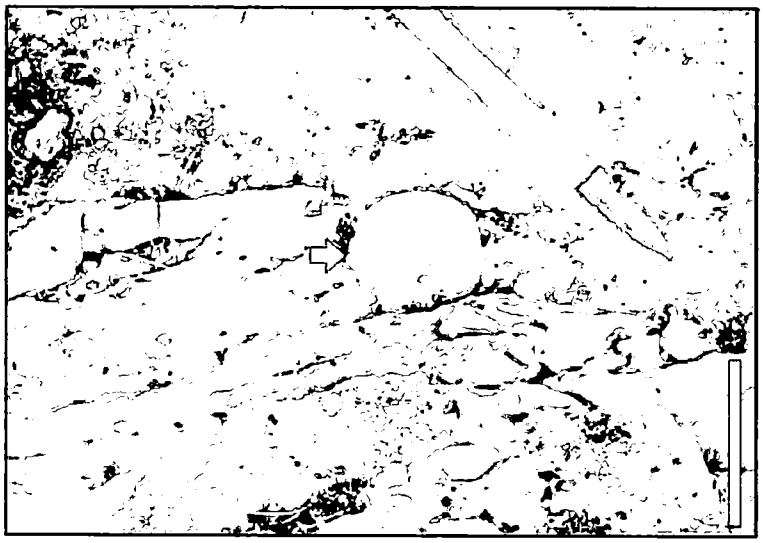


Figure 3.71a *In situ* echinoid preserved within the porcellaneous foraminifera micritic litharenite facies. Scalebar=5cm.

Figure 3.71b. Photomicrograph (PPL) of the porcellaneous foraminifera micritic litharenite facies. Sample LCA 114 (log SA-12, bed 2). (i) miliolid with reddened test, (ii) rounded echinoid plate fragment, (iii) angular siliciclastic grain. Scalebar=2mm.

Figure 3.71c Photomicrograph (PPL) of the porcellaneous foraminifera micritic litharenite facies (sample LCA 139, log SA-40, bed 2). Finer-grained sample, dominated by mud to fine-sand grade siliciclastic grains. Also present is the large encrusting hyaline foraminifera *Planolinderina* (i). Scale bar=2mm.

Planolinderina (0 to 2 %) and small hyaline benthics (0 to 5 %) (**Figures 3.71a** and **b**). Small benthics are dominated by discorbids with rare bolvinids. Flattened, discoidal *Orbitolites* tests are typically fragmented, although rare intact forms are preserved. Tests are up to 7.5 mm. Slightly reddened miliolid tests are up to 2.5 mm and are abraded (**Figure 3.71b**). Other fossils present in this facies include bivalves (0 to 2 %), gastropods (≤0.5 %), fenestrate bryozoa (≤0.5 %) and echinoid spines and plates (0.5 to 2.5 %) (**Figure 3.71b**). Coral fragments are present in some samples (LCA 119 and LCA 120), although this is atypical of this facies in general.

Interpretation: depositional environment

The porcellaneous foraminifera micritic litharenite facies contains a marine faunal and ichnofossil assemblage. Although fragmentation and abrasion of fossils indicates re-working, deposition under normal marine salinities within the photic zone is inferred from the abundance of discoidal *Orbitolites* tests (Leutenegger 1994). As discussed in the above sections, *Orbitolites* may be a free-living or epiphytic form. An abundance of the long-lived larger forms may indicate the presence of perennial vegetation (e.g. sea grasses and coralline algae) (Leutenegger 1984, Hallock and Peebles 1993). Sea floor vegetation may have had a baffling effect, promoting sediment accumulation.

A low-nutrient depositional setting is inferred from the presence of the larger foraminifera *Orbitolites* (Hallock and Glenn 1986, Hallock 1988). However, (localised) high-nutrient areas are inferred from the abundance of burrowing and deposit-feeding organisms such as bivalves, molluscs and gastropods. An abundance of smaller benthic foraminifera is also consistent with high nutrient availability (Travé 1992, Travé *et al.* 1996). Coral colonisation may have been inhibited in this environment by the inferred high nutrient and/or unstable substrates.

Deposition relatively close to the sediment source in a shoreline-proximal marine environment is concluded from the coarse, typically poorly sorted nature of siliciclastics. Limited abrasion due to transport and reworking prior to deposition is inferred from the angular nature of grains. Deposition in a moderate to low-energy environment around fair-weather wave base is concluded from the significant percentage of muddy matrix. Current energies were not strong enough to remove small foraminifera tests.

In summary, the porcellaneous foraminifera micritic litharenite facies represents the deposition of siliciclastic material in a low-energy/partially protected marine environment, close to a source of siliciclastic sediment. In terms of fauna, this facies has similarities to the Porcellaneous Foraminifer Coralline Algae Wackestone Packstone facies of Travé (1992) and Travé *et al.* (1996), interpreted as an inner shelf protected platform. The porcellaneous micritic litharenite described here contains a significantly higher percentage of siliciclastic material. Travé *et al.* (1996) suggest a relative increase in siliciclastic material is a consequence of deposition in increasingly proximal environments closer to a siliciclastic source.

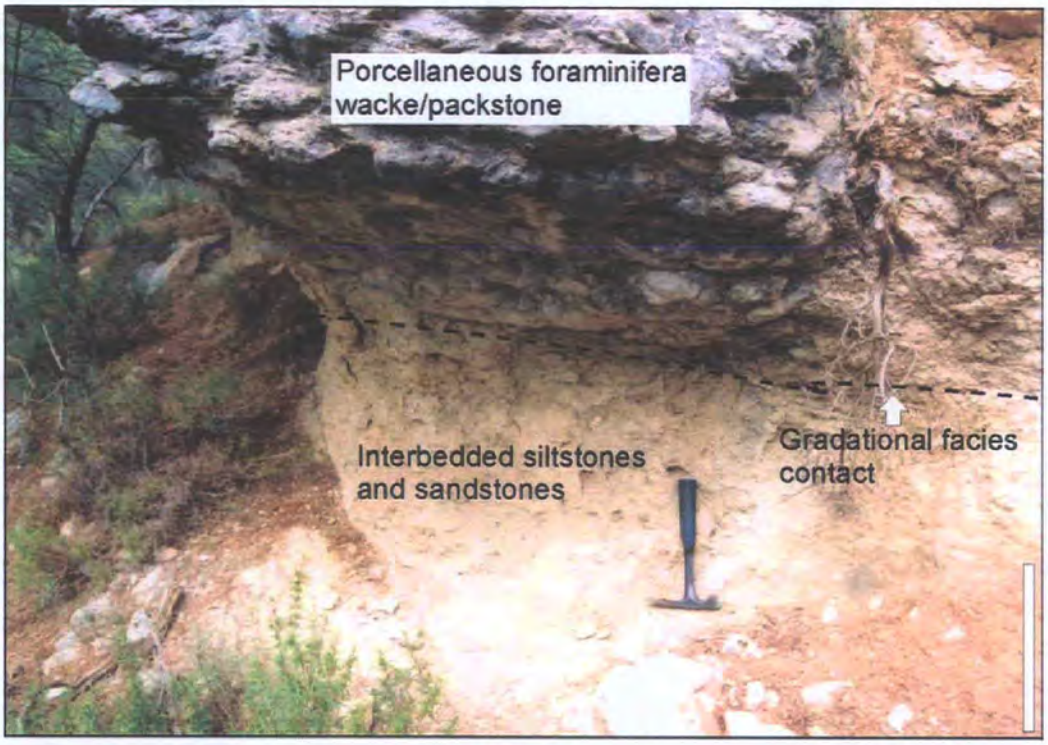
3.5.3.2 Interbedded calcareous litharenitic siltstone and sandstone

Occurrence and bed characteristics

This facies is very common and encountered throughout the Sant Amanc study area (**Figure 3.63**). The interbedded calcareous litharenitic siltstone and sandstone facies is the lowest stratigraphic unit present in the Sant Amanc section, and is always encountered below the porcellaneous foraminifera wacke/packstone and porcellaneous foraminifera micritic litharenite facies (**Figures 3.63** and **3.72a**). This facies can be traced to isolated exposures in fields situated to the south (GR 18352656) and west (GR 17502641-log SA-50, bed 3) of the Sant Amanc ridge (**Figure 3.62**). It is inferred that this facies extends laterally over 1 km, and probably beyond the limits of the study area.

Exposure of this facies is quite poor on the main Sant Amanc ridge section, and true bed thickness cannot be measured here. Measured thickness of this facies is approximately 5 m at GR 17502641. At GR 18352656, exposed thickness is approximately 2.5 m (**Figure 3.73a**). Individual siltstone and sandstone beds are up to 30 cm thick (**Figures 3.73a** and **b**). Upper and lower bed contacts are sharp but non-erosional (**Figure 3.73b**). Cross-stratification of beds is evident in some localities. Beds dip consistently northwest. Beds may also be characterised by the presence of symmetrical ripple and planar lamination occasionally disturbed by vertical and horizontal *Skolithos* and *Thalassanoides*-type burrows. Burrows are up to 10 cm in length. Fresh and weathered surfaces of this facies are a yellow-brown to yellow colour (**Figure 3.73a**).

a)



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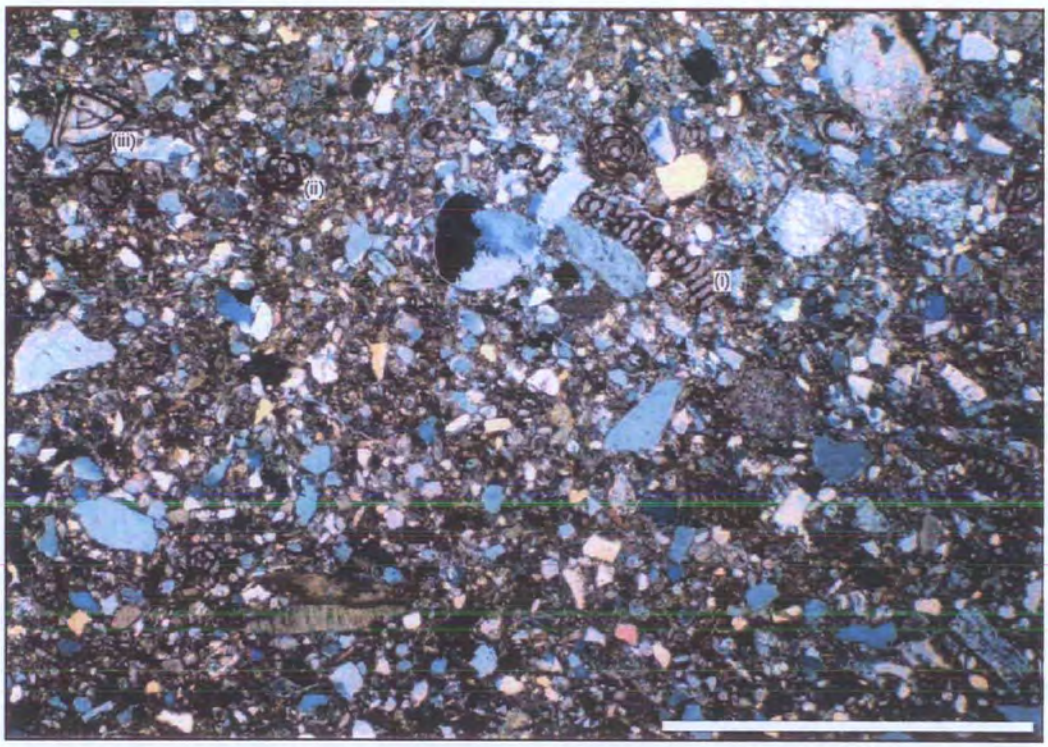


Figure 3.72a Field view of the gradational contact between the interbedded calcareous siltstone and sandstone and porcellaneous foraminifera micritc litharenite lithofacies (log SA-20 beds 1 and 2). Scalebar=50cm.

Figure 3.72b Photomicrograph (XPL) of the interbedded calcareous litharenitic siltstone and sandstone (sample LCA 142, log SA-3 bed 1). Abundant biota include discoidal *Orbitolites* (i), miliolids and *Triloculina* (iii). Scalebar=2 mm

Lithological description

Matrix-supported interbedded sands and silts with very rare bioclasts comprise this facies. Siliciclastic grains are angular, moderately well to poorly-sorted and subspherical. Grainsize ranges from 0.1 to 0.5 mm. Siliciclastic grains identified are detrital quartz (20 to 35.5 %) and lithics (5 to 15 %) with trace amounts of feldspar. Mono- and polycrystalline quartz varieties are present. Lithic grains include siltstone, sandstone and quartz-mica schist. Micas may have a 'rotted', appearance, and are altered to a yellow to rusty-brown coloured clay observed in thin section. Dark-brown micrite matrix comprises up to 51.5 to 64 % of this facies.

Body fossils are quite rare, comprising up to 8 % of this facies. Foraminifera present are *Orbitolites* (0 to 1.5 %), miliolids (0 to 3.5 %) and small hyaline benthics (0.5 %) (**Figure 3.72b**). Flattened, discoidal *Orbitolites* tests are fragmented and abraded. Miliolids (*Quinqueloculina* and *Triloculina*) are preserved intact but are abraded (**Figure 3.27b**). Echinoid spines (0 to 1 %) and undifferentiated skeletal debris (0 to 2 %) are also present.

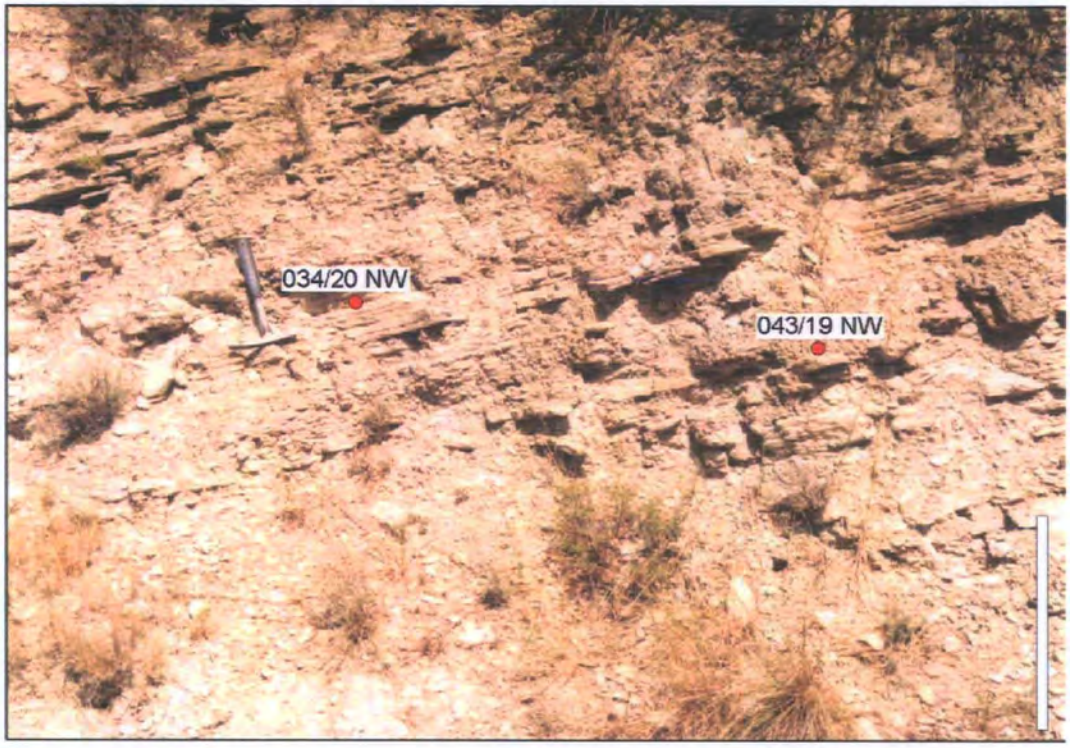
Interpretation: depositional environment

Deposition of this facies in a marine environment is concluded from the moderate diversity of stenohaline fauna. Open marine conditions within the photic zone are implied from the presence of the symbiont-bearing larger foraminifera *Orbitolites*, although tests are not *in situ* and have undergone some transport and abrasion. Nutrient-rich areas are inferred from the presence of burrowing fauna such as echinoids (cf. Travé *et al.* 1996).

Limited abrasion of siliciclastics prior to deposition and lithification is inferred from the angular nature of grains, although the grains are moderately well-sorted. Variable energy conditions are concluded from the presence of fine-grained silt interbeds and the significant amount of micrite matrix. Rapid deposition is inferred by the sharp nature of contacts, although contacts are non-erosional. Deposition at depths above fair weather wave base is inferred from the presence of current ripple lamination and cross-stratification.

In summary, the interbedded calcareous litharenitic siltstone and sandstone facies is interpreted to represent deposition within a shallow marine, low to moderate-energy setting at/above fair weather wave base.

a)



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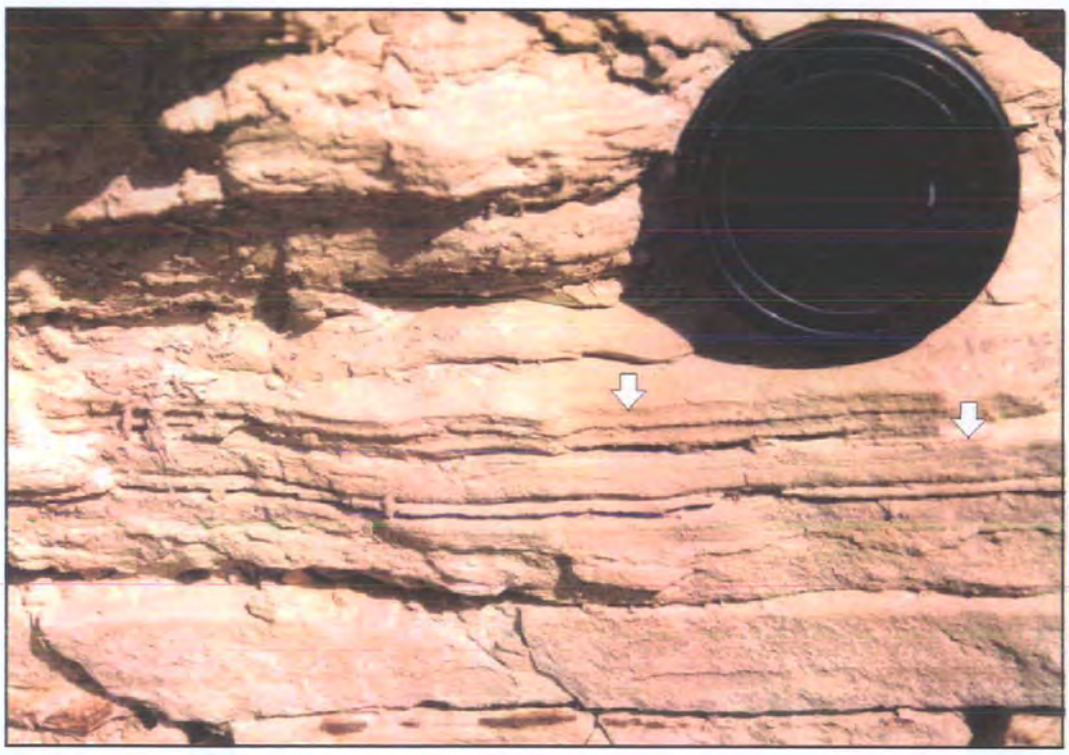


Figure 3.73a Field exposure of the interbedded litharenitic siltstone and sandstone lithofacies, facing north (GR 18352656). This facies is poorly exposed, although at this locality cross-stratification of beds is observed. Measured surfaces yield a north-west progradation direction. Scalebar=1m.

Figure 3.73b Close-up of cross-stratified units. Individual sandbodies have mm/cm-scale parallel laminae (arrowed). Lens cap=6.5cm.

3.6 Sant Amanc depositional model

Four distinct facies have been defined for the Sant Amanc study area, and the characterisitics of each facies are summarised in **Table 3.5**. Facies occur in the same stratigraphic order across the Sant Amanc study area with only minor variations that are discussed below. The typical stratigraphic succession, and an interpretation of the changing depositional environments, is illustrated in **Figure 3.74**.

The first facies to be deposited was the interbedded calcareous litharenitic siltstone and sandstone. This facies can be traced along the entirety of the Sant Amanc ridge, and thicknesses of up to 5 metres are measured in isolated exposures to the south and west (log CA-16, bed 3). Thickness variations cannot be determined across the Sant Amanc area due to poor laterally continuous exposure. This facies is succeeded vertically by the porcellaneous foraminifera micritic litharenite (**Figure 3.74**). The contact is gradational to sharp and planar (**Figure 3.74**). The porcellaneous foraminifera micritic litharenite may be traced along the entire length of the ridge exposure (> 500 m) and is also present present in isolated exposures to the west (log CA-16 bed 4), although this facies grades both vertically (**Figure 3.74**) and laterally (**Figure 3.68**) into the porcellaneous foraminifera wacke/packstone facies. In the western limits of the Sant Amanc area, the porcellaneous foraminifera micritic litharenite is succeeded vertically by the coral wacke/float/bafflestone facies (logs SA-47 and 48). The porcellaneous foraminifera wacke/packstone facies is vertically succeeded by the coral wacke/float/bafflestone facies (**Figure 3.74**). The contact is gradational. The coral wacke/float/bafflestone facies occurs as low relief, mound-like beds that extend laterally for 50 to 100 m (**Figure 3.64**). Beds pinch-out laterally, and the long axis is arranged normal to the palaeoshoreline (**Figure 3.64c**). This facies is vertically succeeded by the porcellaneous foraminifera wacke/packstone facies, which caps the carbonate succession at Sant Amanc (**Figure 3.74**).

The four facies identified within the Sant Amanc sections typically occur together as described above, thus are interpreted to collectively to represent a single mixed carbonate-siliciclastic facies association. Facies contain a variety of stenohaline biota and are interpreted to have accumulated under normal open to slightly restricted marine conditions within the photic zone. The maintainance of moderate to low-energy, possibly partially protected conditions is concluded from the significant percentage of fine-grained matrix present within all facies in the Sant Amanc

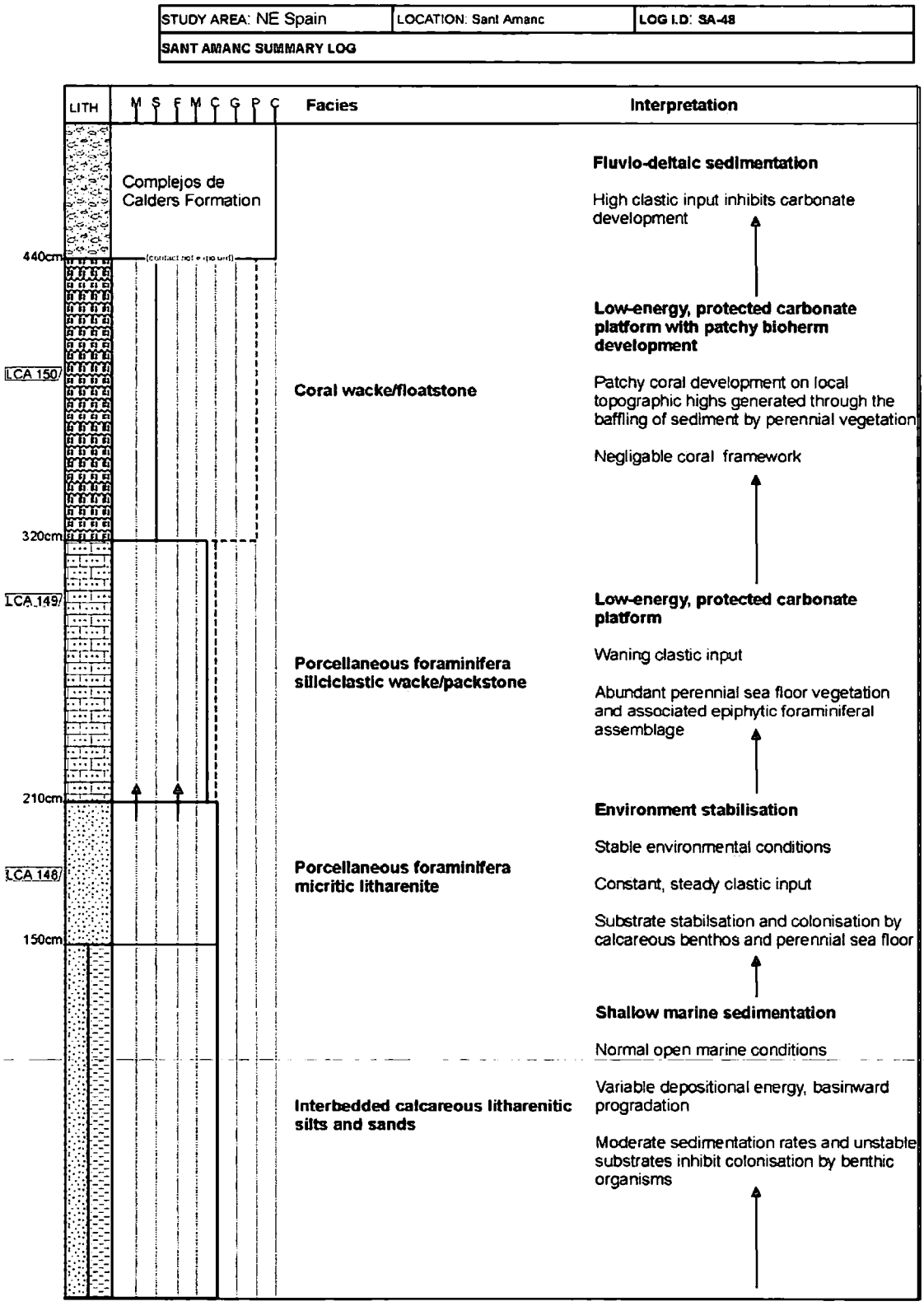


Figure 3.74 Typical log through the Sant Amanc succession.

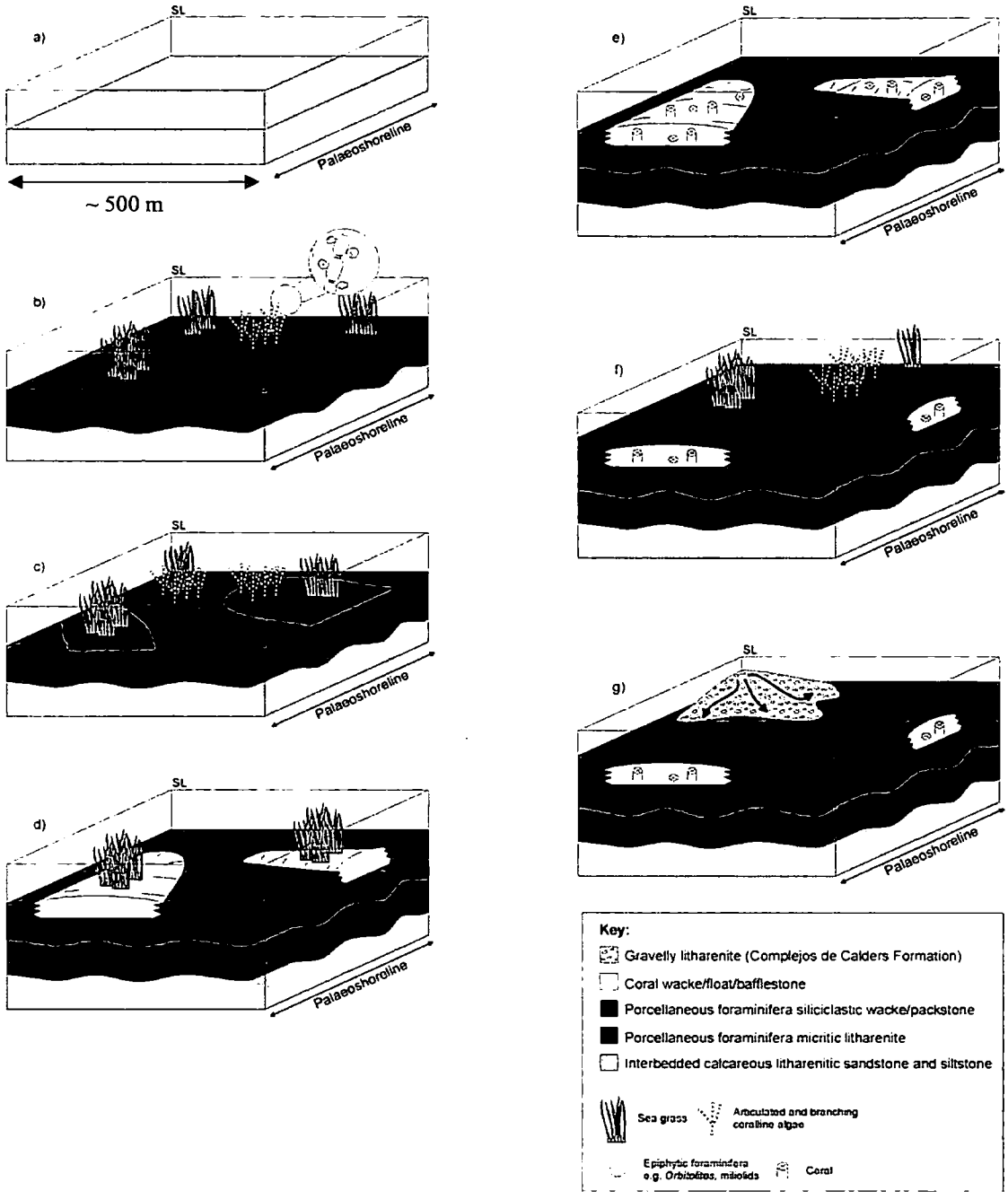


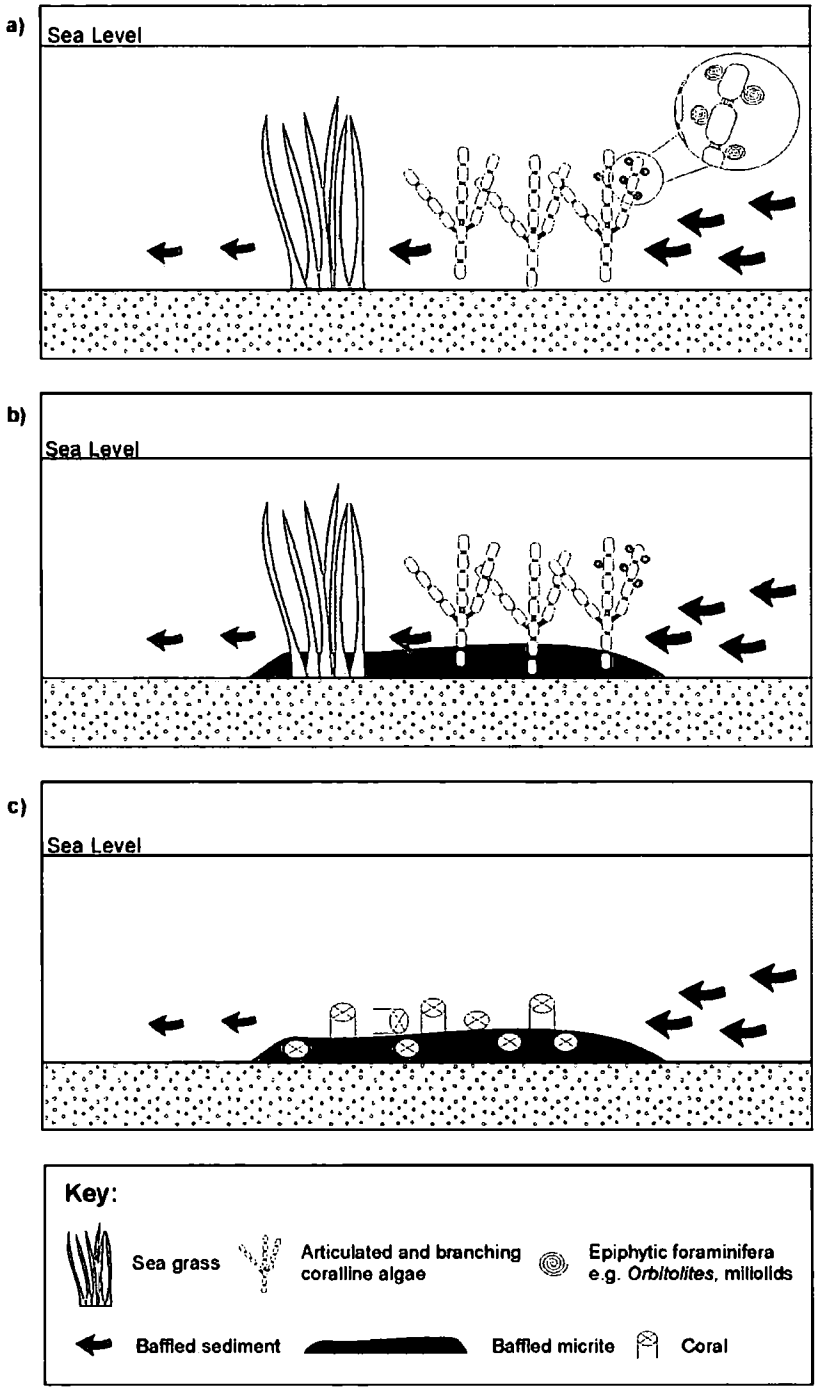
Figure 3.75 Schematic reconstruction illustrating temporal and spatial facies variations in the Sant Amanc study area based upon the correlations illustrated on **Figures 3.63** to **3.65** and **3.68**). The orientation of the palaeoshoreline was roughly east-west. **a)** Widespread deposition of inter-bedded silts and sands. **b)** Decrease in energy and siliciclastic input, tentative colonization by benthic organisms. **c)** Development of muddy, wacke/packstones in low-energy areas. **d)** and **e)** Development of low-relief bioherm. Development may have been aided by the baffling effect of sea floor vegetation. **f)** Progradation of the inner-shelf wacke/packstones and coral development ceases. **g)** Carbonate production is terminated by the input of coarse siliciclastics of the Complejos de Calders Formation. See text for discussion.

succession. The lower 2 to 5 metres of this association are interpreted to represent widespread deposition of cross-stratified sands and silts at water depths at and above fair weather wave base (**Figure 3.75a**). Variable depositional energies are inferred, with sandstone beds representing relatively high-energy clastic inputs and laminated siltstones representing low-energy deposition from suspension. Moderate to low-energy conditions overall are concluded from the absence of erosional structures at the base of sandstones and the abundance of micrite matrix. Variations in coarse clastic input are attributed to the seasonality of the Late Eocene climate in the eastern Pyrenean region (Cavagnetto and Anadón 1996). The measured progradation direction is towards the northwest, roughly perpendicular to the inferred palaeoshoreline orientation. The sedimentary characteristics of this facies are not distinctive of a particular palaeoenvironment, although the biota present, including *Orbitolites,* miliolids and small benthics, are indicative of a protected inner-shelf platform-type setting (Travé 1992, Travé *et al.* 1996, Romero *et al.* 2002).

The transition from interbedded silts and sands to deposition of porcellaneous foraminifera-rich micritic sandstones is interpreted to represent a distinct change in environmental conditions from variable to more stable depositional energies with a constant input of silt to sand grade clastics (**Figure 3.75b**). Stabilisation of the marine substrate and the development of a diverse benthic community under oligotrophic conditions are construed from the increased abundance of larger foraminifera although the presence of nutrient-rich areas is interpreted from the presence of deposits feeding and burrowing organisms and small benthic foraminifera such as discorbids. Colonisation by larger sessile benthic organisms such as corals was inhibited either by the still significant input of siliciclastics, the muddy nature of the substrate or the presence of nutrient-rich areas. It is postulated that the development of perennial sea floor vegetation, (e.g. sea grass, macroalgae) aided colonisation by foraminifera such as *Orbitolites,* and non symbiont-bearing miliolids, through their ability to adapt to an epiphytic mode of life (**Figure 3.75b**).

It is interpreted that the gradational lateral transition of porcellaneous foraminifera-rich sandstones into porcellaneous foraminifera-rich wacke/packstones demonstrates how clastic input/accumulation varied spatially over small distances (less than 20 m). It is postulated that these variations may be due to local currents regimenting siliciclastic transport pathways and/or the presence of a number of

Figure 3.76 Proposed mechanism for the development of low-relief muddy bioherms in the Sant Amanc area.



a) The development of sea-floor vegetation (i.e. sea grasses, articulated coralline algae) local modifies the hydrodynamic conditions forming localized low-energy areas.

b) Sea-floor vegetation effectively baffles particulate sediments and a low-relief mud mound may develop.

c) Even though the relief above the sea floor may only be a few centimeters, these low-relief mounds provide relatively good colonization sites for corals.

localised siliciclastic sources along the shoreline. Significant post-depositional reworking is not consistent with the abundance of intergranular muddy matrix.

The gradual decrease in clastic input over time is seen through the vertical transition from porcellaneous foraminifera-rich sandstones to porcellaneous foraminifera-rich wacke/packstones (**Figure 3.75c**). The decreasing clastic input is accompanied by an increase in biota abundance and diversity. Larger benthic foraminifera, in particular *Orbitolites* and *Rhabdorites,* may have inhabited true benthic as well as epiphytic substrates provided by sea grasses, articulated and branching coralline algae. Although oligotrophic conditions are interpreted from the abundance of larger benthic foraminifera, locally high nutrient levels close to the sea floor are inferred from the abundance of deposit feeding and burrowing organisms, and the abundance of small, non symbiont-bearing benthic foraminifera.

The vertical and lateral gradational transition of the porcellaneous foraminifera wacke/packstone facies into coral wacke/float/bafflestone facies is interpreted to represent the development of low-relief muddy bioherms (**Figure 3.75d**). The presence of perennial sea floor vegetation (sea grasses, articulated coralline algae etc) is inferred from the abundance of the large, long-lived foraminifera *Orbitolites*. It is postulated that the baffling effect of this sea floor vegetation promoted the accumulation of fine-grained sediments leading to the development of localised topographic highs (**Figures 3.75b** and **3.76**). These highs may only have been a few centimetres above the sea floor, but provided suitable sites for coral recruitment. Coral development may have been inhibited within the porcellaneous foraminifera wacke/packstone facies due to a number of factors such as the lack of available substrate, substrate instability or high nutrient levels on the sea floor. Reef development on mobile, sandy substrates along the Jesira and Bajuni Archipelago, Somalia, has been aided through the binding of sediment by the sea grass *Thalassodendon* (Carbone *et al.* 1994). *Thalassodendron* meadows provide support for a diverse community of epiphytes including coralline algae, foraminifera and bryozoa. The sea grass meadows have effectively trapped fine sediment building low flat mounds rising from the surrounding substrate (Carbone *et al.* 1994).

Muddy coral biostromes extended laterally 50 to 100 m in the Sant Amanc area, with the long axis (and thickest part of the accumulation) orientated normal to the inferred palaeoshoreline (**Figure 3.75e**). As a comparison, the long axes of coral

bioherms from the Terminal Complex exposed in the eastern part of the Vic Basin are orientated parallel to the palaeoshoreline as a response to manipulation by shoreline-parallel (longshore) currents (Travé 1992, Travé *et al.* 1996). It is suggested that shoreline-normal currents rather than longshore currents influenced the morphology of the muddy coral bioherms in the Sant Amanc study area. The mechanism for these currents can only be speculated in this thesis due to dearth of palaeoflow data throughout the study area. The degree of current reworking of a structure with low topographic relief would have been limited in a partially protected setting such as that interpreted for the Sant Amanc area. Coral breakage through the action of predators and/or current action is inferred from the presence of coral fragments in laterally equivalent wacke/packstones. Corals are loosely packed, with rudstone and bafflestone textures only locally developed. It is thought that the lack of framework development, attributed to limiting environmental conditions (e.g. turbidity, high nutrients), prevented coral biostromes developing into bioherms.

A further porcellaneous foraminifera wacke/packstone unit, interpreted to represent progradation of the protected carbonate inner-shelf, may succeed the coral wacke/float/bafflestone facies (**Figure 3.75f**). The upper contact between the coral bioherms and the overlying wacke/packstones is gradational and non-erosional, thus rates of coral growth and carbonate production were insufficient to build up to sea level.

The cessation of carbonate development in the Sant Amanc area is attributed to siliciclastic input (**Figure 3.75g**). Trough cross-stratified pebbly sandstones, interpreted as part of the fluvio-deltaic Complejos de Calders Formation, followed by red silty mudrocks of the Artès Formation, vertically succeed the carbonate succession. Travé (1992), and Travé *et al.* (1996) conclude that the top of the Terminal Complex in the eastern part of the Vic Basin is an unconformity. This conclusion cannot be verified in the Sant Amanc area as the contact is poorly exposed.

3.7 Summary

The Calders and Sant Amanc sedimentary successions were deposited within an over all siliciclastic-dominated setting with no significant slope (inferred from the gentle inclination of beds and absence of slump or slide deposits). Where colonised by

carbonate producers, the areas were dominated by a benthos of coralgal-foraminifera reefs and diverse larger benthic foraminifera (the palaeoecology of which are described in **Appendix 2**), which were used to define facies (**Sections 3.2 and 3.4**).

The Calders sedimentary succession represents the development of foraminifera and coralline algal shoals on abandoned siliciclastic substrates. Waning siliciclastic influx was typically marked with the colonisation of stabilised substrates by larger benthic foraminifera and development of laterally extensive foralgal shoals. Localised coral growth occurred within the shoals in relatively seaward environments. The change from carbonate development to siliciclastic deposition was rapid. The mechanism invoked for such a sharp, laterally extensive facies change may be switching of delta lobes.

The mixed carbonate-clastic succession exposed in the Sant Amanc area is interpreted to be an as yet unstudied exposure of the Terminal Complex, as defined by Travé (1992), and Travé *et al.* (1996). These sediments were deposited within a partially protected, inner-shelf, shallow marine setting at depths around and above fair weather wave base within the photic zone. Carbonate production and diversification of the biota was influenced by silciclastic input. Oligotrophic conditions prevailed in the water column, although mesotrophic, and possibly eutrophic conditions, existed at the sediment-water interface. These high-nutrient areas supported a diverse community of deposit feeding and burrowing organisms, as well as small benthic foraminifera. Sea floor vegetation, a mixture of sea grasses with branching and articulated coralline algae, acted as a baffle, effectively trapping fine-grained sediment and encouraging the development of low-relief coral bioherms. Carbonate development was eventually terminated through massive clastic input as the Pyrenean Foreland Basin moved from its under-filled to over-filled stage.

In conclusion, the Calders and Sant Amanc depositional successions demonstrate the variability of facies that may be encountered on shallow/marginal marine shelf systems. The successions demonstrate that facies vary laterally over small distances (10s to 1000s metres). This is comparable to facies changes observed in modern reef environments. These lateral and vertical facies variations are attributed to changes in siliciclastic input, relative water depth and depositional energy.

4. Regional Geology and Field Areas, South-Eastern Spain

4.1 Introduction

The aim of this chapter is to summarise previous work on the regional geological and palaeoenvironmental evolution of the Betic Cordillera in order to put the studied section in the Fortuna Basin into a regional context. Particular attention will be paid to the post-Cenozoic evolution and development of the Neogene intramontane basins, of which the Fortuna Basin is an example. Although this chapter aims to summarize the geology of the region, it should be noted that the Neogene tectonic evolution of the Betic-Alboran-Rif region is still controversial (Gracés *et al.* 2001).

4.2 Regional geology of the Betic-Alboran-Rif Area

The geology of the southwestern Mediterranean region has been influenced by two major tectonic events. These are the oblique collision of the Iberian microplate with the African plate (and closure of the Tethys Ocean) during the Cretaceous to Miocene, and major extensional events in the Oligo-Miocene (Dewey *et al.* 1989, Sanz de Galdeano 1990, 1992; Doglioni *et al.* 1997, Calvert *et al.* 2000).

The Betic Cordillera, together with the Rif (northern Africa), is the western-most part of the circum Alpine-Mediterranean orogenic chain that evolved through collision of the Iberian and African plates during the Cenozoic (Sanz de Galdeano and Vera 1992, Lonergan and White 1997). Neogene continental deformation along the African-Iberian plate boundary was distributed over a broad zone 500 km wide, extending from the Betic Cordillera (SE Spain) to the High Atlas mountains (northern Morocco) (**Figure 4.1**) (Calvert *et al.* 2000). The NE-SW striking Betic and Rif mountain belts form an arcuate orogen, delimiting the Miocene to Recent Alboran Basin (**Figure 4.1**) (Calvert *et al.* 2000, Doglioni *et al.* 1997, Lonergan and White 1997). Continental collisional was accommodated by over 200 km of north-south shortening during the Late-Creataceous to Early Miocene, and 50 km of northwest-southeast shortening from the Late Miocene onward (Soria *et al.* 2000, Watts *et al.* 1993, Dewey *et al.* 1989, Platt and Vissers 1989). In SE Spain, this shortening was accompanied by the development of the Guadalquivir Foreland Basin (**Figure 4.1**) (Fernandez *et al.* 1998, Lonergan and White 1997).

The later stages of orogenesis in the Middle to Late Miocene (Tortonian) throughout the western Mediterranean region were characterised by coeval shortening and extension accommodated by detachment processes and strike-slip movement (Montenat *et al.* 1992, Sanz de Galdeano and Vera 1992, Doglioni *et al.* 1997, Lonergan and White 1997, Martinez-Martinez *et al.* 1997). A compressional tectonic regime with contemporaneous shoshonitic to lamphroitic volcanism, dominated the Betic region from the Tortonian onward. (Montenat *et al.* 1992, Martinez-Martinez *et al.* 1997). The Betic Cordillera underwent general uplift from the late Miocene to Pliocene (Sanz de Galdeano and Vera 1992).

4.2.1 Tectonic Provinces of the Betic Cordillera

The Betic Cordillera occupies a belt in southern Spain extending 600 km east to west and up to 200 km north to south (**Figure 4.1**) (Soria *et al.* 1999, Sanz de Galdeano and Vera 1992). The Cordillera is not a continuous mountain range, consisting of several mountainous areas with intervening Miocene basins (**Section 4.2.1.4**). The Betics consist of four major tectonic provinces that were clearly delimited by the early Neogene (Balanya *et al.* 1987). These are 1. External Zones, 2. Internal Zones, 3. the Campo de Gibraltar Complex and 4. Neogene extensional basins (**Figure 4.1**) (Calvert *et al.* 2000, Lonergan and White 1997, Sanz de Galdeano and Vera 1992 and others). The characteristics of each tectonic province will be summarised in the following sections.

4.2.1.1 The Internal Zones

The allochthonous syn-orogenic sediments of the Betic Internal Zones were originally deposited on the Alboran sub-plate (original Tethyan oceanic crust) that was located to the east of the present Betic Cordillera (Calvert *et al.* 2000, Lonergan and White 1997, Sanz de Galdeano and Vera 1992). During the Oligocene, the Internal Zones of the Betics and Rif, as well as the Tertiary metamorphic belts of the Tell Atlas, Sardinia, Calabria, Sicily and Kabylies, constituted a NE-SW trending

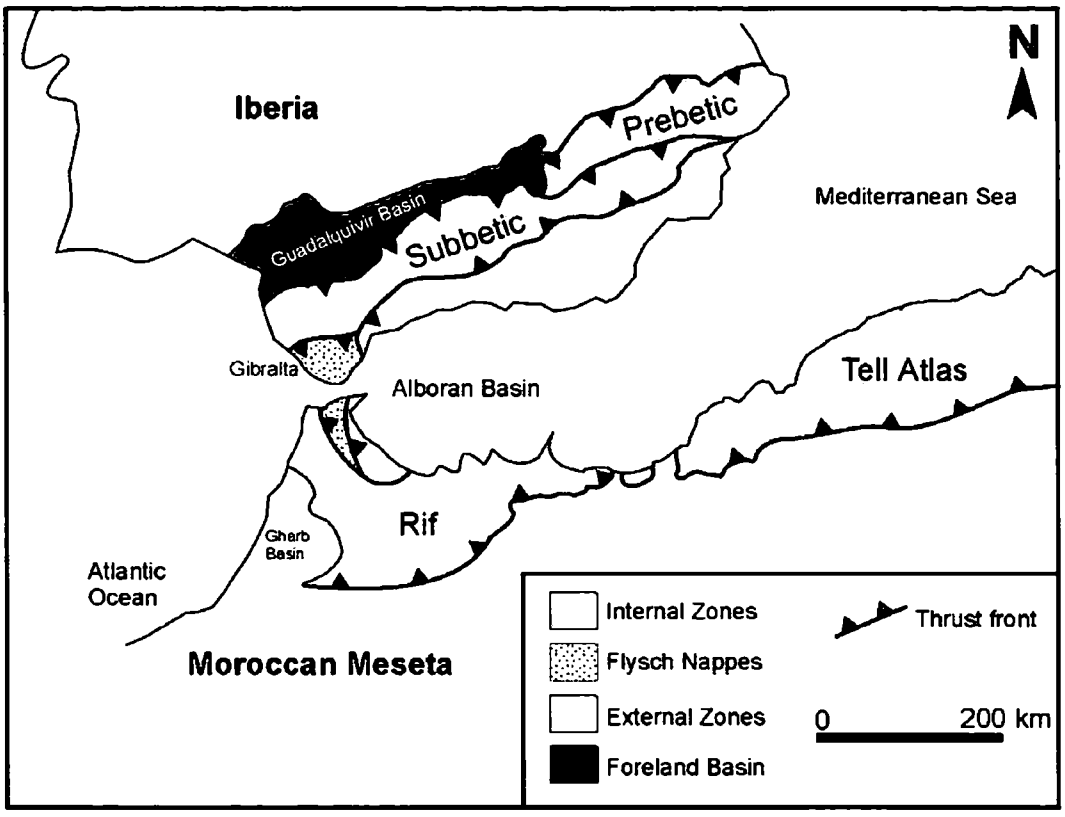


Figure 4.1 Schematic sketch of the main tectonic provinces of the western Mediterranean region. Modified from Calvert *et al.* (2000).

Cretaceous-Paleogene mountain belt that existed along the coasts of southeastern France and eastern Spain (**Figure 4.2a**) (Calvert 2000, Lonergan and White 1997). This mountain belt was situated to the northeast of a northward dipping subduction zone (Lonergan and White 1997). Large extensional basins, (including the Alboran Basin) evolved in the western Mediterranean during the Late Oligocene to Early Miocene as a consequence of rollback of this subducting slab (Jolivett and Faccenna 2000, Zeck 1999, Vergés and Sabat 1999, Doglioni *et al.* 1997, Lonergan and White 1997). Slab rollback led to the southward migration of the subduction zone and extension and dispersal of the Internal Zones (**Figure 4.2b**) (Jolivet and Faccenna 2000, Vergés and Sabat 1999, Lonergan and White 1997, Coward and Dietrich 1989). When the subduction zone collided with the North African continental margin, it separated into two major fragments. An eastern fragment continued to roll back and eventually evolved into the Calabrian Arc forming the Tyhrrenian Sea (**Figure 4.2b**). The western fragment rolled back to the west, generating the Alboran Sea and the Betic-Rif orogen as the Internal Zones were emplaced onto the Iberian and North African passive margins during the Early to Middle Miocene (**Figure 4.2b**) (Lonergan

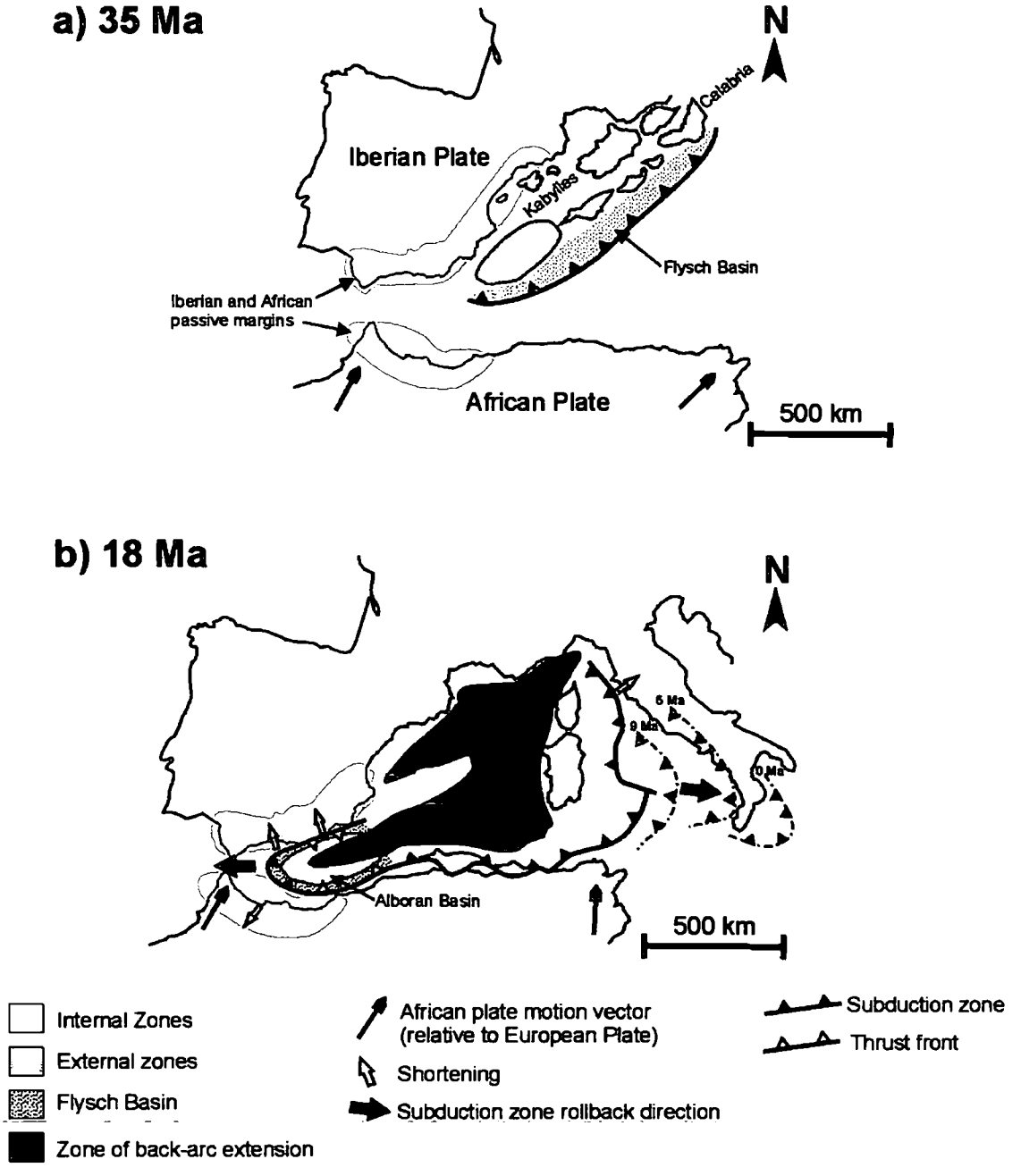


Figure 4.2 Major geodynamic events in the western Mediterranean. (a) During the Late Oligocene (~35 Ma), Internal Zone rocks were located behind a northward-dipping subduction zone. Flysch deposits of the Campo de Gibralta complex may also have been deposited here. (b) In the mid-Miocene (~18 Ma), the subduction zone split to form two branches that roll back to the east and west. Internal Zone rocks were emplaced onto the Iberic and African plate margins. After Lonergan and White (1997).

and White 1997). The Internal Zones were emplaced onto the southern passive margin of the Iberian Massif as nappe structures, with the load inducing subsidence of the Iberian margin and the compressive deformation of the Betics (Banks and Warburton 1991, van der Beek and Cloetingh 1992).

The Internal Zones of the Betic Cordillera contain three tectonically superposed nappe complexes that are exposed as east-west trending domes: the Nevada-Filabride Complex, the Alpujarride Complex and the Malaguide Complex (Sanz de Galdeano and Vera 1992). The original boundaries between these complexes have been cut or reactivated by large low angle extensional detachment faults (Lonergan and Platt 1995, Platt and Vissers 1989, Platt 1986). The Nevada-Filabride and Alpujarride Complexes consist of Precambrian, Palaeozoic and Triassic metasediments. These rocks have been affected by early high-pressure, low-temperature metamorphism associated with crustal stacking in the Betics, and later low-pressure high- temperature metamorphism related to extension and exhumation of deep crustal rocks (Balanyá *et al.* 1997, Monié *et al.* 1994, Bakker *et al.* 1991, De Jong 1991, Goffé *et al.* 1989). The Malaguide Complex contains relatively unmetamorphosed Palaeozoic, Mesozoic and Cenozoic sediments (Sanz de Galdeano and Vera 1992).

4.2.1.2 The External Zones

The Betic External Zones are the folded and detached sedimentary cover that was originally deposited on the submerged southern Iberian and African plate margins during the Mesozoic to Cenozoic (**Figure 4.2a**) (Sanz de Galdeano and Vera 1992, Fernandez *et al.* 1998, Comas and García Dueñas 1988). The sedimentary cover is divided into the Subbetic and Prebetic zones (**Figure 4.1**). The Subbetics are composed of deep marine sediments including fine-grained turbidites (the 'Tap' deposits) with olistostromes (Sanz de Galdeano and Vera 1992). The Prebetics are predominantly composed of siliciclastics with minor algal carbonates that were deposited in a coastal, shallow marine environment (Sanz de Galdeano and Vera 1992).

Major WNW-ESE orientated compression affected the Subbetics during the late Aquitanian-Burdigalian as a consequence of the westward drift of the Internal Zones (**Figure 4.2a**) (Lonergan and White 1997, Sanz de Galdeano and Vera 1992).

The consequence of this was deformation and the development of internal erosional unconformities. The Prebetics were affected to a lesser extent (Sanz de Galdeano and Vera 1992).

4.2.1.3 The Campo de Gibralta Complex

The allochthonous flysch units of the Campo de Gibralta Complex are exposed in the Gibralta region in Spain, with time-equivalent sediments exposed in the Tell region of Morocco (**Figure 4.1**). Sediments are turbidites and hemipelagites originally deposited during the Mesozoic to Miocene on Tethyan crust in the deep subduction trough to the north of of the Internal Zones (**Figure 4.2a**) (Lonergan and White 1997). Fragments of the Campo de Gibralta Complex were relocated onto the southern Iberian passive margin with the Internal Zones during the lowermost Miocene (**Figure 4.2a**) (Lonergan and White 1997).

4.2.1.4 The Neogene Basins

Neogene sedimentation in the Betic region occurred within the recently developed Guadalquivir Foreland Basin (**Figure 4.1**), and a number of strike-slip basins that developed predominantly within the Internal Zone massifs (**Figure 4.3**). Basins in southeastern Spain (including the Fortuna Basin within which the studied section at Altorreal is situated) developed along the Internal-External Zone boundary within a wide (>250 km) left-lateral shear zone (the Trans-Alboran Shear Zone) trending NE-SW from Alicante to Almeria (**Figure 4.5**) (Sanz de Galdeano and Vera 1992, Montenat and Ott d'Estevou 1990, Montenat *et al.* 1987).

The Neogene strike-slip basins contain a record of Neogene and Quaternary syn- and post-orogenic sediments that lie uncomformably on Betic basement. The different basins contain remarkably similar sedimentary facies, including pelagic marls, fanglomerates, turbidites, reefal carbonates and evaporites (Soria *et al.* 1999, Lonergan and Schreiber 1993, Pomar 1991, Braga *et al.* 1990, Martin *et al.* 1989, Santisteban and Taberner 1988). These inter-linked basins are thought to have evolved within in the Betic Strait (also referred to as the Guadalentin Corridor) (Sanz de Galdeano and Vera 1992). The Betic Strait was an open marine connection between the Atlantic Ocean and the Mediterranean Sea that existed from earliest Tortonian to early Messinian time (**Figure 4.4**) (Sanz de Galdeano and Vera 1992).

Surface uplift of the central Cordillera caused relative lowering of sea level and basins lost their marine character during the late Tortonian-Messinian (Sanz de Galdeano and Vera 1992).

4.3 Study Area: The Fortuna Basin

Formation of the Fortuna Basin was initiated during the early Tortonian as a faulted basin (Gracés *et al.* 2001, Sanz de Galdeano and Vera 1992). Movement along major faults (the Crevillente and Alhama de Murcia Faults, which approximately mark the present northern and southern basin margins respectively) controlled basin evolution (**Figures 4.5** and **4.6**) (Gracés *et al.* 2001, Poisson and Lukowski 1996). The basin-fill is predominantly siliciclastic with isolated intercalations of carbonates and evaporites (Montenant 1973, Gracés *et al.* 2001, Krijgsman 2000, Poisson and Lukowski 1996, Santisteban and Taberner 1988). In northern areas, sediments were sourced mainly from External Zone massifs that consisted predominantly of Mesozoic-Cenozoic carbonates (Montenat 1973). Siliciclastic input from the Internal Zones was restricted to alluvial-fan delta systems in southern basin areas (Montenat 1973). Relicts of the Internal Zone rocks on the southern basin margins include the Orihuela and Callosa Massifs to the SE, and the Sierra Espuña mountain range in the SW (**Figures 4.6** and **4.7**) (Gracés *et al.* 2001).

The Fortuna Basin existed in the Tortonian-Messinian as a 15 to 20 km wide, 60 km long trough (Gracés *et al.* 2001). The basin-fill succession consists of three basin-wide units: the Lower Marine Unit, the Evaporite Unit and the Continental Unit.

4.3.1 Early Tortonian- the Lower Marine Unit

The Fortuna Basin had an open marine connection to the adjacent strike-slip basins during the early Tortonian through the Betic Strait (Montenat *et al.* 1990). During this time, subsidence rates greatly exceeded the rate of sedimentation. This resulted in a significant relative deepening of the basin and deposition of thick accumulations of pelagic sediments (Gracés *et al.* 2001). The Lower Marine Unit is up to 500 m thick, consisting of turbidites and pelagic marls collectively known as the Los Baños Formation (known also as the Fortuna Marls) (Müller and Hsü 1987). The marls grade laterally into deltaic and reef facies (Lonergran and Schrieber 1993 and

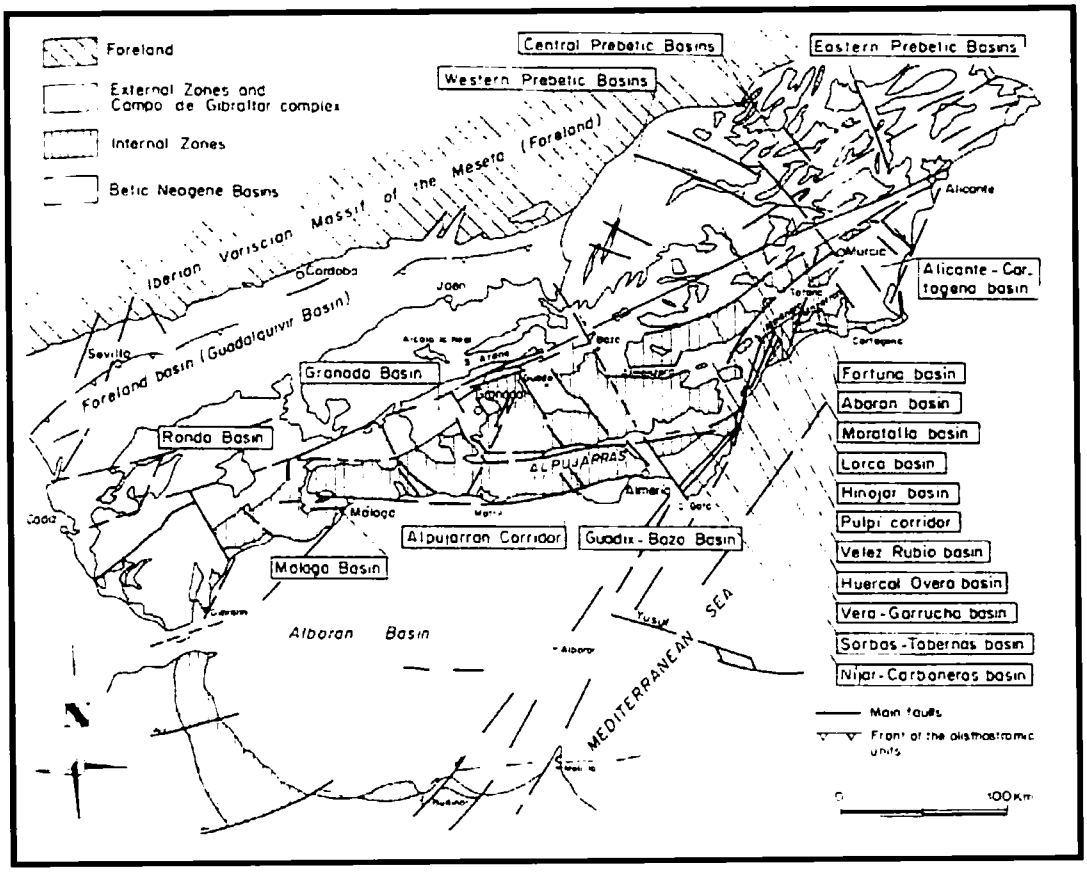


Figure 4.3 Location of important Miocene intramontane basins of the Betic Cordillera. After Sanz de Galdeano and Vera 1992

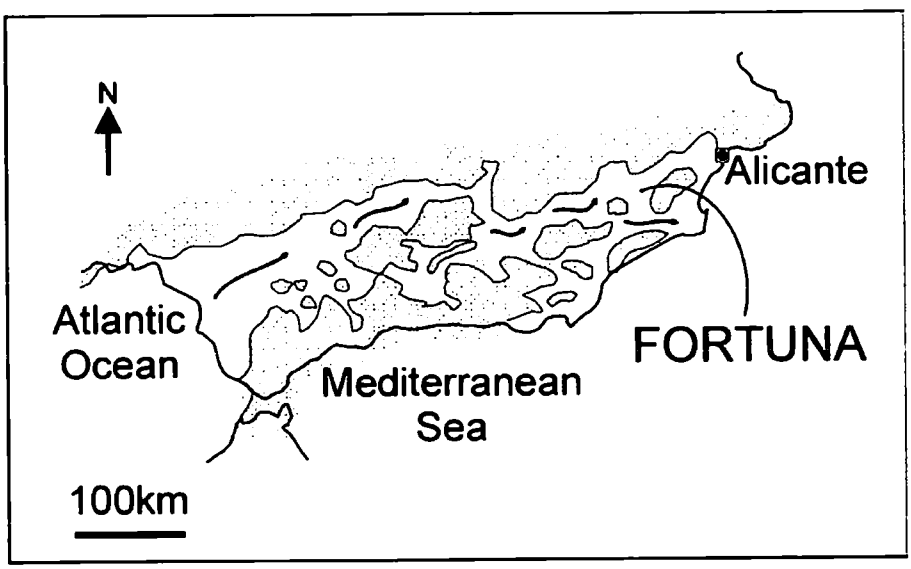


Figure 4.4 Neogene open sedimentation in the Betics occurred within the Betic Strait- a corridor connecting the Mediterranean with the Atlantic Ocean. Tectonism lead to closure of the strait in the late Tortonian, leading to basin restriction. The approximate location of the Fortuna Basin is indicated. Shaded areas correspond to emergent areas during the early to late Tortonian. Modified from Reinhold (1995).

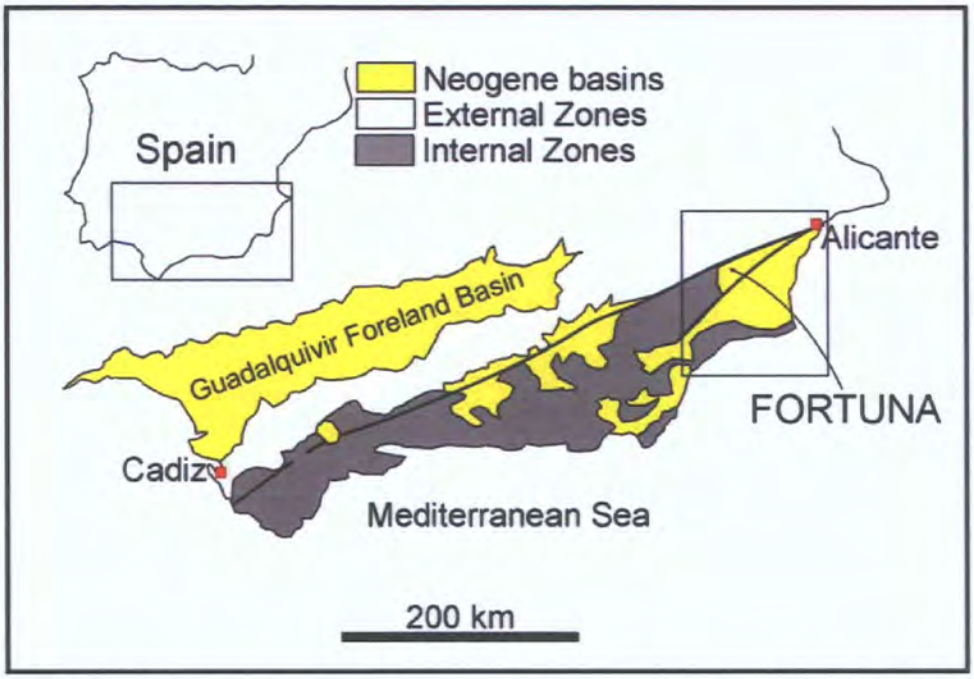


Figure 4.5a Summary map of the geology of the Betic Cordillera region, SE Spain. The Fortuna Basin is indicated. Redrawn from Garcés *et al.* (1998)

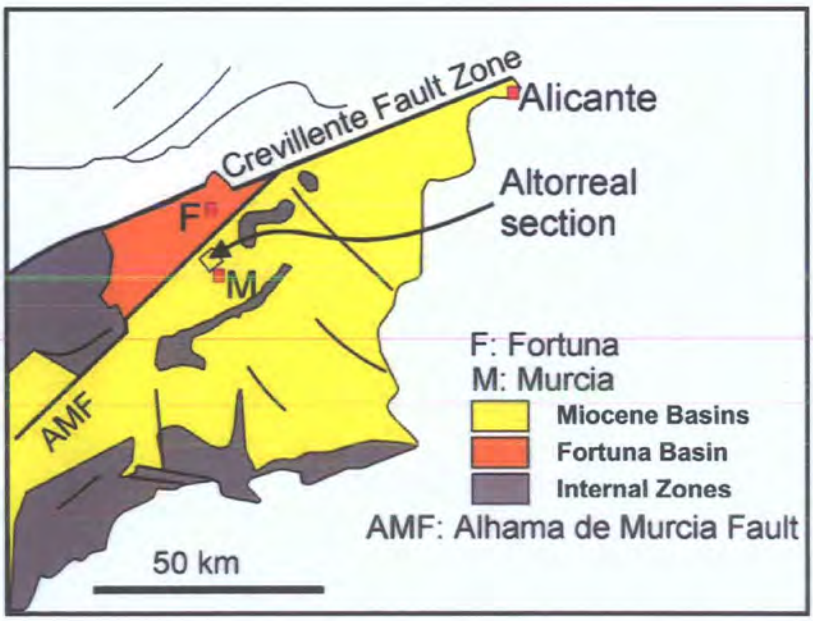


Figure 4.5b Detail of above, indicating the location of the studied section at Altorreal, in the Fortuna Basin. **F**= Fortuna, **M**=Murcia, **AMF**= Alhama de Murcia fault. The present limits of the Fortuna Basin are highlighted in orange. In the Tortonian, the basin extended further south just to the north of Murcia. Redrawn from Garcés *et al.* (1998)

Santisteban 1981). These interdigitating deltaic and reefal facies were studied during this project. Significant thickness variation of the pelagic units across the basin at this time is related to a complex underlying horst and graben structure (Gracés *et al.* 2001).

4.3.2 Late Tortonian-the Evaporite Unit

A drastic change in basin configuration occurred during late Tortonian time (Gracés *et al.* 2001). Restriction of marine water circulation and the initiation of hypersaline conditions are related to tectonic actitivity along the Alhama de Murcia fault and the evolution of a topographic high in the east (Gracés *et al.* 2001). The resultant Evaporite Unit is up to 200 m thick and consists of a regressive succession of gypsiferous marls, diatomites, massive gypsum and coarse terrigenous siliciclastics collectively known as the Rio Chicamo Formation (Müller and Hsü 1987).

Restriction of the Fortuna Basin is coeval with increasing sedimentation rates that were up to 1.0 m kyr^{-1} (Gracés *et al.* 2001). The relative shallowing trend of the upper Tortonian sedimentary succession represents the rapid basin filling as sedimentation rates exceeded the rate of creation of accommodation space through subsidence (Gracés *et al.* 2001). Once the Fortuna Basin reached the 'overfilled stage', sedimentation rates decreased to 30-40 cm kyr^{-1}, matching the generation of accommodation space (Gracés *et al.* 2001). Movement of major fault systems in the region manifest themselves through the development of numerous angular uncomformities (Montenat and Ott d'Estevou 1990).

4.3.3 Late Tortonian-Messinian-the Continental Unit

During the Late Tortonian-Messinian, the Fortuna Basin was partially confined with localised shallow lacustrine and palustrine environments in distal and marginal areas of large alluvial fan systems (Gracés *et al.* 2001). Basin confinement led to development of evaporites (the Rio Chicamo Formation) that grade vertically into continental sediments of the Rambla Salada Formation (Müller and Hsü 1987). The Rambla Salada Formation is characterised by lacustrine grey marl and limestones alternating with alluvial red clays and conglomerates (Müller and Hsü 1987). Relative sea level fall associated with the Mediterranean Messinian Salinity Crisis 5.5 to 5 Ma

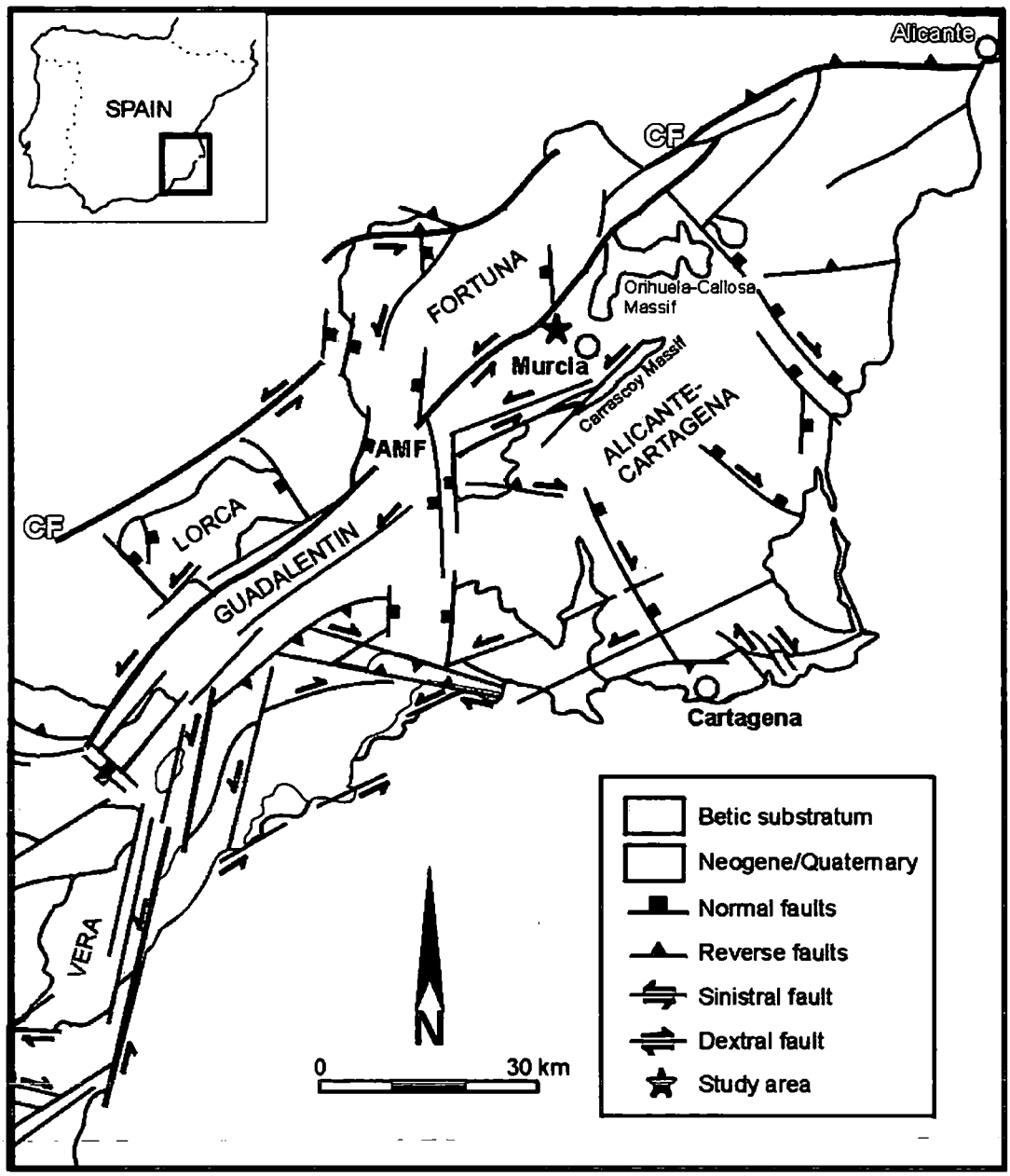


Figure 4.6 Structural map of the eastern Betic region. The evolution of the Fortuna Basin was influenced by movement aliong the Crevellient Fault (**CF**) and the Alhama de Murcia Fault (**AMF**). Modified from Krijgsman *et al.* (2000).

led to a significant decrease in accommodation space in the Fortuna Basin. Alluvial systems rapidly prograded basinward, depositing coarse siliciclastic sediments in distal marine trough environments (Gracés *et al.* 2001, Krijgsman *et al.* 1999). Although largely confined, the Fortuna Basin maintained a minor open marine connection to the south-southeast during the Messinian (Müller and Hsü 1987).

4.3.4 Late Messinian-Pliocene

A marine transgression and the (temporary) re-establishment of basin-wide normal marine conditions in the Fortuna Basin occurred during the latest Messinian to Pliocene (Gracés *et al.* 2001). The marine influx was confined to narrow depressed areas parallel to the NW-SE striking Alhama de Murcia fault (Gracés *et al.* 2001).

The final stage of basin evolution was controlled by a phase of transpressional left-lateral shear with N-S to NNW-SSE compression along the Alhama de Murcia fault during the Pliocene (Gracés *et al.* 2001). The consequence was uplift of the eastern basin margin-the present Carrascoy Massif (**Figure 4.6**) (Sanz de Galdeano 1998). Syn-depositional folding of Pliocene sediments was related to basin uplift (Gracés *et al.* 2001).

4.3.5 The Tortonian Salinity Crisis

Several intramontane marine basins in the eastern Betics contain a thick late Miocene evaporite unit. Evaporites of the inner Neogene basins (those located furthest from the Mediterranean during the late Miocene, including the Fortuna and adjacent Lorca Basins) have been dated as upper Tortonian to lower Messinian (Krijgsman *et al.* 2000, Dinares-Turell *et al.* 1999). Marginal basins (those situated in the vicinity of the Mediterranean, including the San Miguel de Salinas, Almeria, Sorbas and Nijar-Carboneras Basins) contain evaporite units dated as late Messinian to early Pliocene. It is these younger deposits that have been associated with the famous Mediterranean 'Messinian Salinity Crisis' (Playà *et al.* 2000). The evaporites of the inner basins typically have a mixed marine and non-marine character (Krijgsman *et al.*, 2000). In contrast, the evaporites of the marginal basins have a marine signature throughout (Playà *et al.* 2000).

Widespread marine sedimentation in the Fortuna Basin ended in the Late Tortonian (7.8 Ma) (Krijgsman 2000). Interpretation of the Evaporite Unit has proved controversial, mainly due to the difficulty of dating evaporitic strata and the lack of biostratigraphic control (Garcés *et al.* 2001). Müller and Hsü (1987), Lukowski *et al.* (1988) and Dinares-Turell *et al.* (1999) have related the evaporites of the Fortuna Basin to the Mediterranean Messinian Salinity Crisis. In contrast, Krijgsman (2000) relates the evaporites to a localised phase of basin restriction in the Tortonian named the Tortonian Salinity Crisis (TSC). The duration of basin restriction is dated by Krijgsman (2000) as 200 ka (7.8 to 7.6 Ma). The TSC was followed by continentalisation of most of the basin 7.6 Ma as a consequence of the regional tectonic uplift of the metamorphic complexes along the eastern basin margin (Garcés *et al.* 2001, Krijgsman *et al.* 2000).

4.3.6 Lateral Correlation with other Neogene basins

The same broad open marine to restricted evaporitic-diatomitic and continental sedimentary succession observed in the Fortuna Basin is also present to the west in the adjacent Lorca Basin (**Figure 4.5**) (Garcés *et al.* 2001). Magnetostratigraphic correlation has shown that the Fortuna and Lorca Basins had a similar Miocene evolution. Both basins were influenced by movement along the Alhama de Murcia and Crevelliente Faults (Garcés *et al.* 2001). Continentalisation of the Fortuna and Lorca Basins 7.6 Ma was roughly coeval with a regression event in the marginal Guadix-Baza and Granada Basins in the central and western Betics respectively (Soria *et al.* 1988).

4.3.7 Study Area-the Altorreal Section

The study area in SE Spain is covered by the 1:50 000 Orihuela geological map published by the Instituto Tecnológico Geominero de Espana (ITGE) (Boer *et al.* 1974). The equivalent 1:50 000 Orihuela topographic map (Sheet 19, 27-36), published by the Servicie Geográfico Ejército, was also used. An additional 1:20 000 Santomera topographic map, published by the Instituto Geográfico Nacional, was used.

The sedimentary succession studied in the Fortuna Basin is located approximately 10.0 km north of the city of Murcia (and the A-7 motorway), 15.0 km south of Fortuna and 9.0 km to the west of the Sierra de Orihuela, in the southern limits of the basin. The location of the Altorreal section is illustrated on **Figures 4.6** and **4.7**. Sediments are exposed within an east-west trending quarried valley system approximately 500 m in length. Isolated outcrops are also exposed 0.7 to 1.0 km south of the main valley. The maximum stratigraphic thickness is 50 m. The lithological unit exposed in the Altorreal section is the Los Baños Formation.

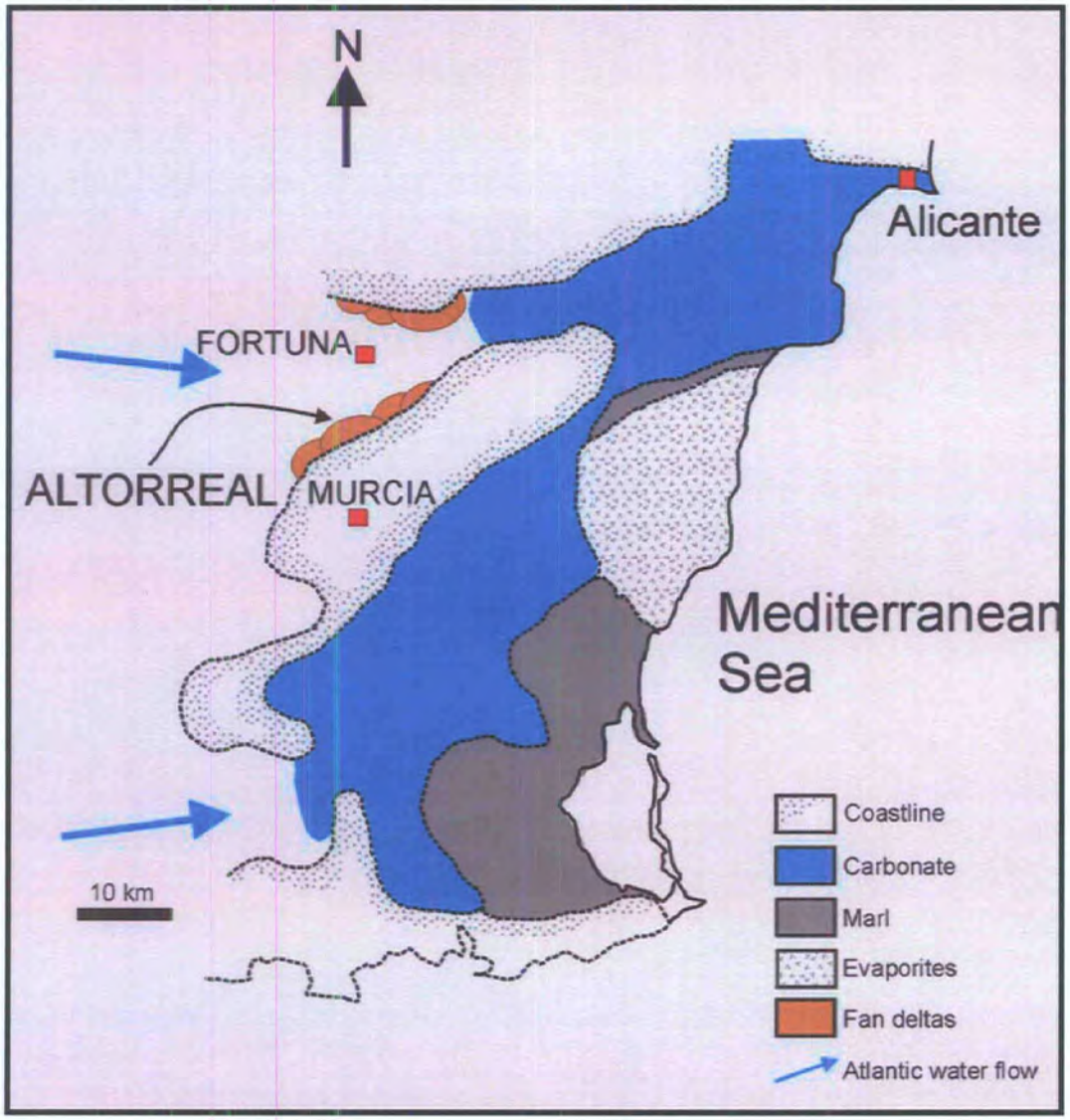


Figure 4.7 Palaeogeographic map of the Betic Strait in SE Spain during the Tortonian. The studied section at Altorreal was situated in the southern sector of the Fortuna Basin. Fan delta systems on the southern margins were supplied with detritus from Internal Zone massifs that were topographically positive features at this time. Modified from Reinhold (1995).

5. Altorreal facies and palaeoenvironmental analysis

5.1 Introduction

The principal aims of Chapter 5 are to describe and synthesize facies data from the Altorreal study area and to present a three-dimensional sedimentological and palaeoecological model. This section briefly summarises the regional geology and previous work undertaken in the immediate area, described in detail in **Chapter 4**. A description and environmental interpretation of all facies present in the Altorreal study area are described in **Section 5.2.** Although not the focus of the study, **Section 5.3** briefly discusses diagenetic features and porosity evolution. Facies associations, and the palaeoenvironmental evolution of the area are described in **Section 5.4**.

Many authors have described the sedimentology of the Miocene intramontane basins of SE Spain. Poisson and Lukowski (1990), Sanz de Galdeano and Vera (1990), Pomar (1991), Braga and Martin (1996) and Soria *et al.* (1999) discuss the basins in terms of sedimentary evolution in response to the eustatic and tectonic evolution of the region. Santisteban and Taberner (1988), Martin *et al.* (1989), Braga *et al.* (1990), Riding *et al.* (1991), Lonergan and Schreiber (1993), Rheinhold (1995) and Wrobel and Michalzik (1999) have discussed the relationship between shallow marine carbonate and siliciclastic facies in selected basins. In addition, significant work has been published concerning basin restriction and the Mediterranean Messinian Salinity Crisis (Gracés *et al.* 1998, Dinarès-Turell 1999, Krijgsman *et al.* 2000, Playá 2000). A detailed facies scheme of the studied Altorreal section within the Fortuna Basin has not until now been undertaken, although Santisteban (1981) and Santisteban and Taberner (1988) have produced a brief review of coral species and the relationship between corals and siliciclastics.

Miocene sediments of the Los Baños Formation are the objects of this study. Metasediments of the Betic Internal Zones (Fortuna Basin basement rocks) are not exposed in the Altorreal section, although basement is exposed to the Sierra Espuña to the northeast and Carrascoy Massif to the south (**Figure 4.4**). The logging, sampling and petrological techniques used in this study are described in **Appendix 1**. For thin section data refer to **Appendix 4. Appendix 6** presents logged sections that encompass field and thin section information. Comparisons are made with relevant

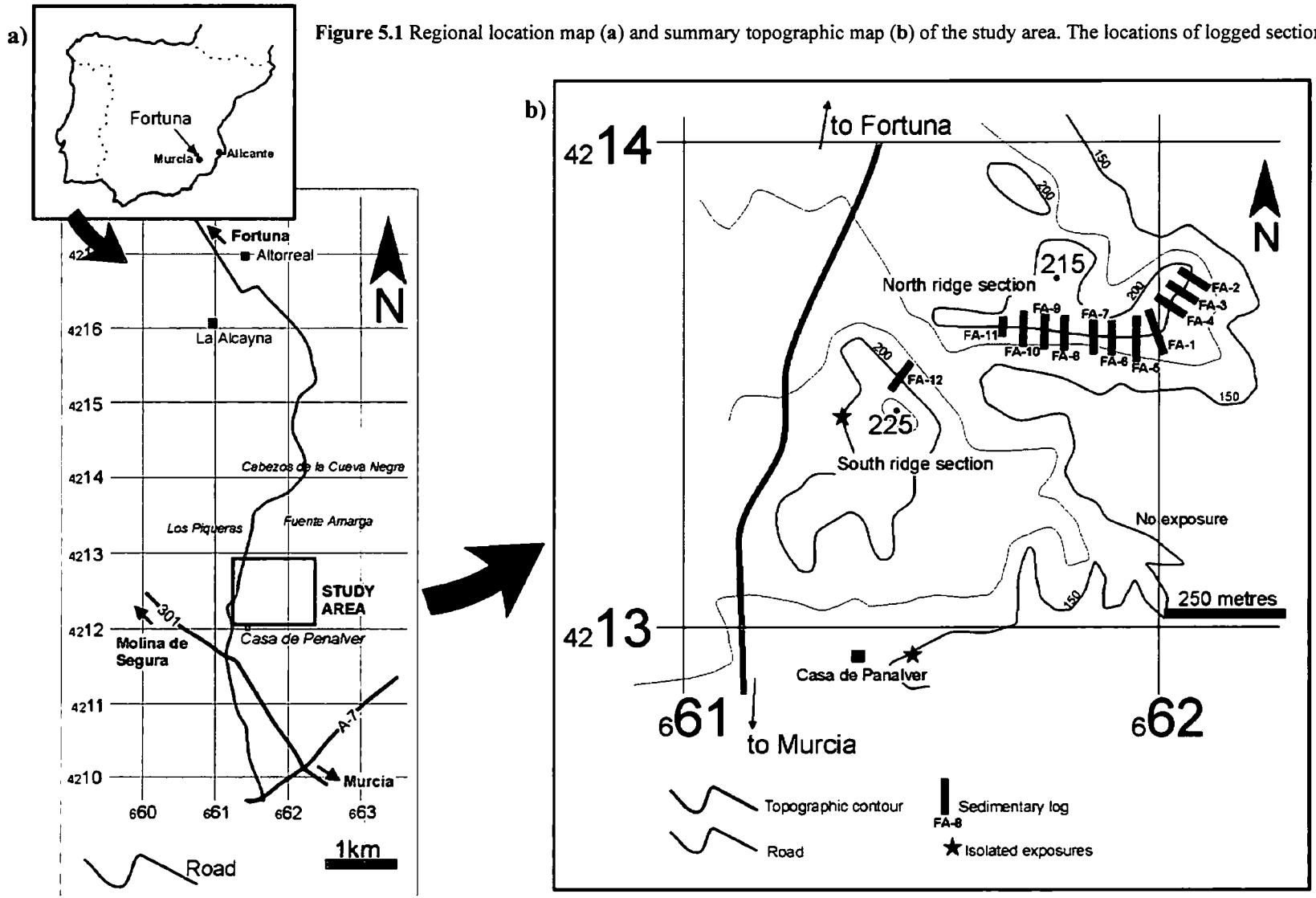


Figure 5.1 Regional location map (**a**) and summary topographic map (**b**) of the study area. The locations of logged sections are indicated.

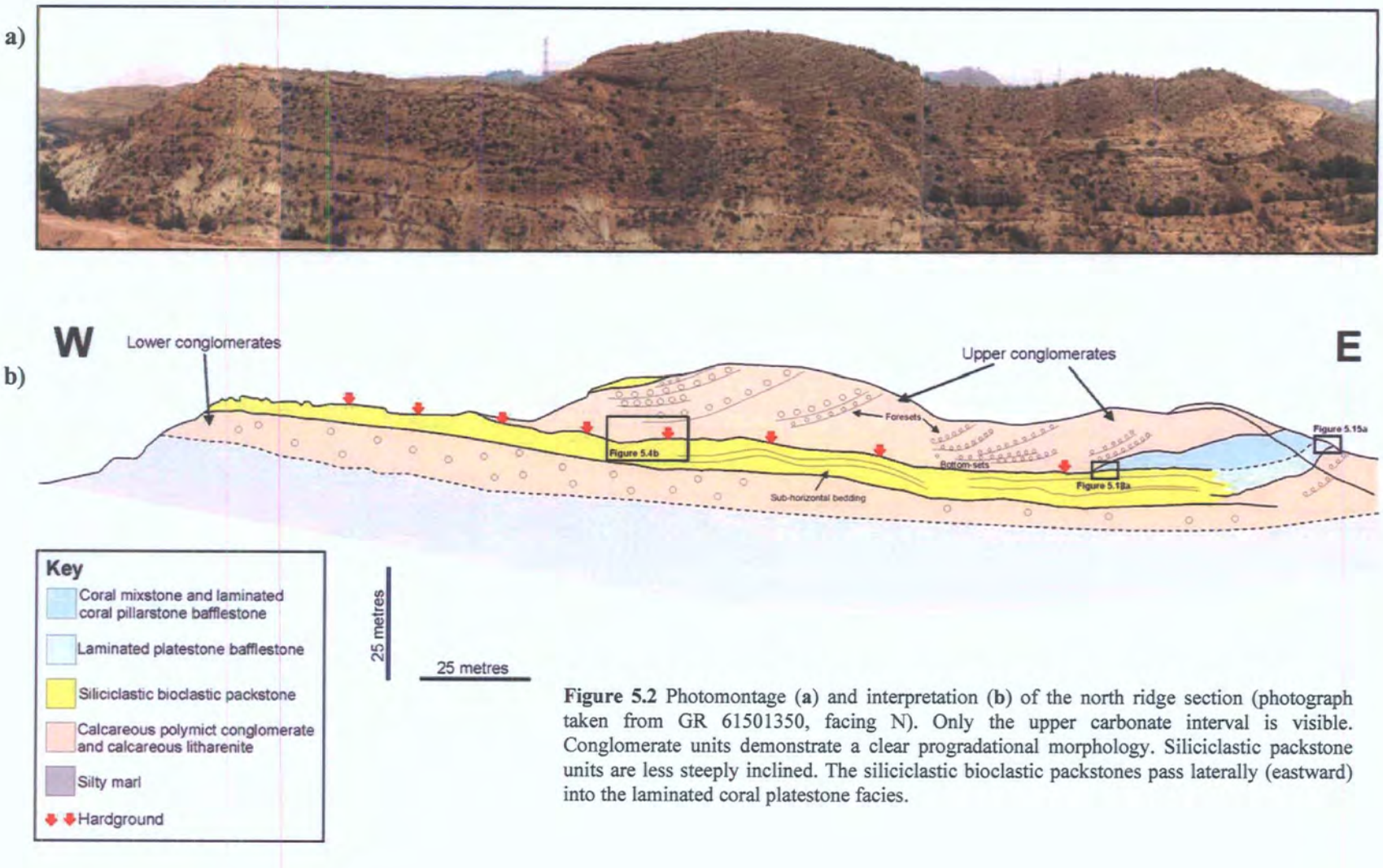


Figure 5.2 Photomontage (**a**) and interpretation (**b**) of the north ridge section (photograph taken from GR 61501350, facing N). Only the upper carbonate interval is visible. Conglomerate units demonstrate a clear progradational morphology. Siliciclastic packstone units are less steeply inclined. The siliciclastic bioclastic packstones pass laterally (eastward) into the laminated coral platestone facies.

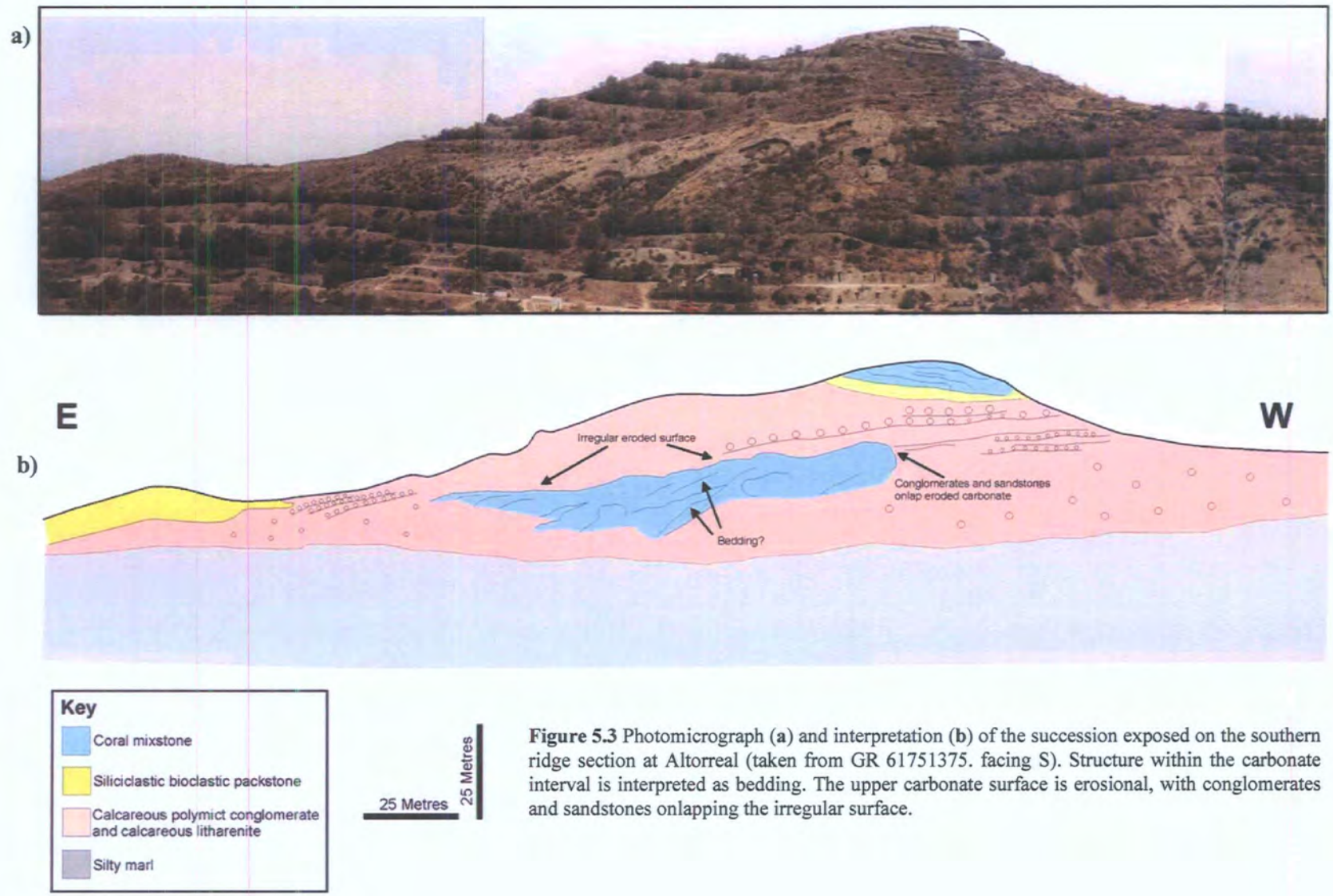


Figure 5.3 Photomicrograph (**a**) and interpretation (**b**) of the succession exposed on the southern ridge section at Altorreal (taken from GR 61751375. facing S). Structure within the carbonate interval is interpreted as bedding. The upper carbonate surface is erosional, with conglomerates and sandstones onlapping the irregular surface.

Table 5.1 Facies scheme for the Alforreal study area, SE Spain

Facies group and facies	Dominant textures/lithologies	Bed thickness	Bedding characteristics	Bed contacts	Lateral continuity	Components	Sedimentary structures	Energy level	Environmental interpretation
Siliciclastic-dominated facies (**Section 5.2.1**)									
Calcareous lithoclastic conglomerate	Conglomerate Conglomeratic sandstone	10 to 50 cm	Conglomerate beds are interbedded with sandstones and siltstones. Metre-scale cross-stratification. Set height up to 20 m.	Upper sharp and non-erosional. Lower sharp and non-erosional to erosional.	> 200 m	Angular to well-rounded lithic grains up to boulder-grade supported within a coarse sand matrix. Minor bioclasts include robust bivalves, planktonic foraminifera and echinoids. Rare *in situ Porites* coral colonies.	Beds may be massive or show reverse grading. Scours.	High to low	**Prograding fan delta lobe.** Deposition of sediments from ephemeral cohesive mass flows and surging debris flows. Colonization by an opportunistic biota between flow events.
Calcareous litharenite	Sandstone Pebbly sandsonte	Tabular units: 0.3 to 2 m. Lenticular units: <0.3 m	Weathers to a brown-yellow colour. Beds are often associated with the calcareous lithoclastic conglomerate facies.	Upper and lower sharp and often erosional.	Tabular beds: > 100 m Lenticular beds: Up to 2 m	Subangular to rounded quartz, metasediments and limestone. Minor bioclasts include oysters, echinoids, gastropods, planktonic foraminifera, coralline algae and *Amphistegina*	Tabular beds are largely massive, sparse wave ripples and parallel lamination Lenticular units demonstrate trough cross-stratification.	Low to moderate	**Shallow marine lower shoreface sands** Deposition within relatively low-energy, delta-front to prodelta environments
Mixed carbonate-siliciclastic facies (**Section 5.2.2**)									
Silty marl	Marl Silty marl	Uncon-strained	Weathers to white mud.	Lower contact not exposed. Upper contact erosional.	> 200 m (beyond the limits of the study area)	Predominantly micrite with subordinate planktonic foraminifera, echinoids and silt-grade, angular quartz	Bedding-parallel mm-scale laminae	Very low	**Low-energy open marine** Deposition below fair weather wave base. Prodelta-setting

Table 5.1 (cont.)

Facies group and facies	Dominant textures/lithologies	Bed thickness	Bedding characteristics	Bed contacts	Lateral continuity	Components	Sedimentary structures	Energy level	Environmental interpretation
Mixed carbonate-siliciclastic facies (**Section 5.2.2**)									
Siliciclastic bioclastic packstone	Packstone Wacke/packstone Pack/grainstone	0.4 to 3.3 m	Beds weather to a pale brown/buff colour. Upper surface of beds may be extensively bored by *Lithophaga* or encrusted by coralline algae. Thin peloidal interbeds.	Lower contacts gradational over 50 cm. Upper contacts are sharp and erosional.	> 200 m	Dominated by molluscs (mainly oysters, pectens and gastropods) with coralline algae, bryozoa, serpulids, planktonic and benthic foraminifera.	Well-bedded, although beds amalgamate along section from east to west.	Moderate	**Open marine carbonate development** Periodic winnowing by currents. Moderate input of detritus from adjacent high-energy areas. Hardground development.
Carbonate-dominated facies (**Section 5.2.3**)									
Laminated coral platestone	Platestone Sheetstone	< 1 m	Beds weather to a pale-grey to white colour. Beds are characterized by bedding-parallel irregular laminae.	Gradational upper and lower.	20 to 25 m	Alternating dark and light-brown micrite with silt-grade siliciclastics. Wacke/packstone matrix locally preserved.	Irregular laminae laterally extensive > 1 m	Moderate	**Shallow marine sub-tidal setting** Microbialite development
Coral mixstone	Mixstone	< 10 m	Wacke/ packstone matrix has weathered out. Beds develop as laterally restricted clinoforms.	Lower sharp, undular and non-erosive. Upper contact erosive. Depth of erosion up to 1 m.	~50 m	Corals, bivalves, gastropods, coralline algae, echinoids, *Amphistegina*, victoriellids and rare silt-grade siliciclastic grains.	None.	Moderate to low	***In situ* coral framework development** Normal marine conditions. Degradation of environmental conditions led to development of columnar strmatolites.

Table 5.1 (cont.)

Facies group and facies	Dominant textures/lithologies	Bed thickness	Bedding characteristics	Bed contacts	Lateral continuity	Components	Sedimentary structures	Energy level	Environmental interpretation
Carbonate-dominated facies (**Section 5.2.3**)									
Laminated coral pillarstone	Pillarstone	2 m	Facies is only encountered at the lateral margins of carbonate units where units pinch out. Beds are characterized by leached corals and weather to a pale brown to pale grey colour.	Gradational lower contact. Erosional upper.	~ 5 m	Stick-like vertical branching corals with microbial crusts, echinoids, bivalves, bryozoa, coralline algae and planktonic foraminifera.	N/A	Moderate to high	***In situ* branching coral development within a shallow marine setting** Contemporaneous microbial growth Significant coarse-grained siliciclastic input during the later stages of branching coral development.

facies examples, specifically those from the neighbouring Neogene Basins of SE Spain and the Mediterranean (**Figure 4.2**).

5.2 Altorreal facies

Carbonate lithologies are classified using the schemes of Dunham (1962) and Insalaco (1998). Classification schemes are presented in **Appendix 1**. Facies classification is based upon a combination of field and thin section data. A detailed facies scheme for the Altorreal area is presented in **Table 5.1**. Facies are classified according to a number of lithological, palaeontological and diagenetic characteristics and occur in three distinct facies groupings: 1) predominantly siliciclastic, 2) mixed carbonate siliciclastic and 3) a predominantly carbonate grouping (**Table 5.1**).

A summary map of the study area and photomontage of the northern and southern Altorreal ridge sections are illustrated on **Figures 5.1 to 5.3** respectively.

5.2.1 Lithoclast-dominated facies

These facies are predominantly siliciclastic in composition although they contain a moderate carbonate component as lithoclasts, cement fragments, bioclasts or cement.

5.2.1.1 Calcareous lithoclastic conglomerate

Lithologies: Calcareous lithoclastic conglomerate
Pebbly calcarenite
Siltstone

Occurrence and bed characteristics

The calcareous lithoclastic conglomerate facies is volumetrically the most important facies in the Altorreal area, and is encountered in all logged sections (**Figures 5.2 and 5.3**). The calcareous lithoclastic conglomerate can occur in association with any of the facies described in the studied area. The lower contact with the marly silt facies is erosional (**Figure 5.4a**). The depth of erosion is up to 50

cm. The calcareous lithoclastic conglomerate overlies and may grade down-dip into the calcareous litharenite facies (**Figure 5.4b**).

This facies is well-bedded with measured bed thickness ranging from 10 to 50 cm. Upper and lower bed contacts are sharp. Lower contacts may be erosional. This facies is composed of conglomerate beds intercalated with thin (< 10 cm) pebbly siltstone and calcarenite beds. Beds are arranged to form cross-stratified sigmoidal clinoform geometries up to 20 m thick and extending laterally over 200 m (**Figure 5.2**). Clinoforms have a progradational geometry. The apparent dip of individual beds increases vertically within a clinoform package (from 2° to around 16°) (**Figure 5.2**). In three-dimensions, large-scale trough cross-stratification of beds is observed (**Figure 5.4b**). Individual beds may be characterised by coarsening upward trends and localised clast imbrication (**Figure 5.5a**).

Lithological description

Exposed beds of this facies weather to a red-brown colour. Fresh surfaces are a pale red-brown colour. The calcareous lithoclastic conglomerate facies is characterised by very poorly sorted granule/pebble to cobble grade lithic clasts with a calcarenitic coarse sand grade matrix. Measured clast size ranges from 0.25 to 75 cm. Pebble to cobble-grade clasts are dominated by a variety of low-grade metasediments, that include quartzite (0 to 17 %), marble (0 to 8.5 %), and laminated quartz-mica schist (21 to 37.5 %), with laminated limestone (**Figure 5.5b**) (4.5 to 41.5 %), microconglomerates (0 to 29.5 %) and intraclasts (**Figure 5.7**). Intraclasts are marl (1.1 to 35.5 %), siliciclastic packstone (0 to 28.5 %) and laminated coral pillarstone (0 to 4.5 %). Marl clasts are typically stained red, pale green or yellow. Angular laminated coral pillarstone boulders up to 75 cm in diameter occur on discrete bedding planes and have been bored by lithophagid bivalves (**Figure 5.5c**). There are no significant changes in grain composition along the studied section (**Figure 5.6**). Clast morphology is highly variable with sub-spherical and sub-angular to well rounded clasts.

Pebble to boulder-grade clasts are supported within a poorly sorted silt to granule-grade matrix. Matrix comprises 5 to 60 % of this facies. Grain compositions of the matrix reflect conglomeratic clasts. Lithic grains, that include quartz-mica schist (2 to 19 %) and laminated limestone (0 to 10 %), are well rounded, poorly

a)

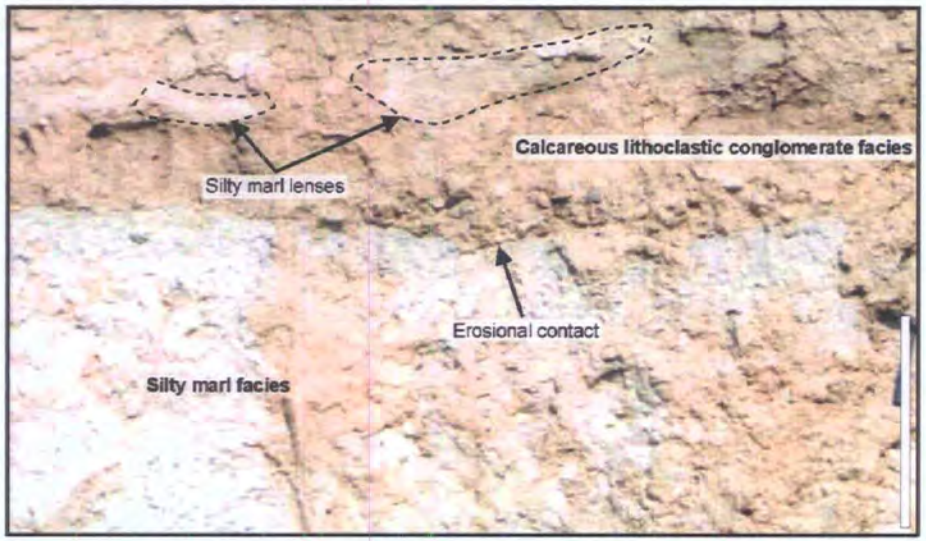


Figure 5.4a Field photograph (GR62501350, facing E) demonstrating the erosional lower contact of the calcareous lithoclastic conglomerate with the silty marl facies. Scalebar = 1 m. The depth of erosion is less than 50 cm. Lenses of silty marl occur in the lower 2 m of the conglomerate.

Figure 5.4b The stratigraphic relationship between the calcareous lithoclastic conglomerate, calcareous litharenite and siliciclastic bioclastic packstone facies. Conglomerates demonstrate metre-scale trough cross-stratification. The angle of foresets increases vertically within a conglomerate package. Two distinct erosively based conglomerate packages are identified. These are interpreted as two different phases of delta lobe progradation. The progradation direction (blue arrows) is obliquely into the page (roughly north).

b)

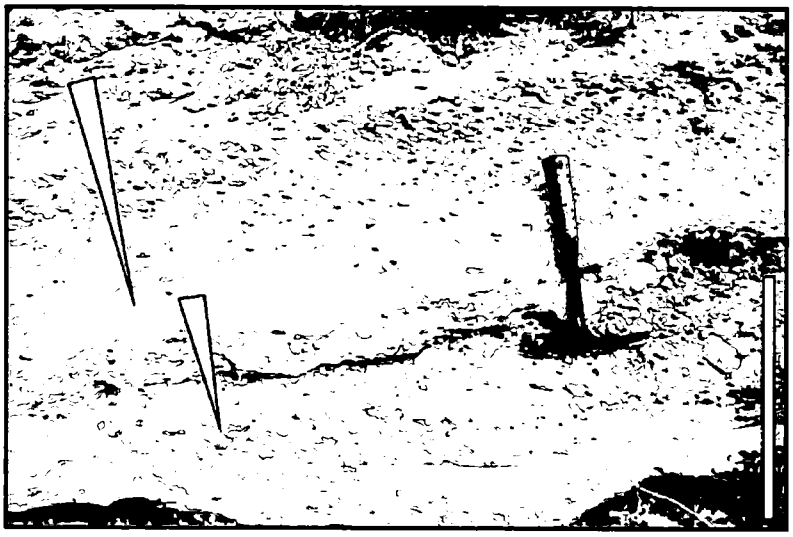


Figure 5.5a Coarsening upward sequences (indicated) within relatively sand-rich beds of the calcareous lithoclastic conglomerate facies (log CA-5, bed 6). Scalebar = 50 cm.

Figure 5.5b Out-sized limestone boulder within the calcareous lithoclastic conglomerate facies. Note the abundant cylindrical borings on the upper surface (arrowed) (log CA-9, bed 10). Scalebar = 50 cm.

Figure 5.5c Out-sized laminated coral pillarstone boulder within the calcareous lithoclastic conglomerate facies (log CA-9, bed 10). It is interpreted that this is a large intraclast derived from the neighbouring area of coral development. Scalebar = 50 cm.

sorted and sub-spherical (**Figure 5.7b**). Lithic grainsize ranges from 0.05 to 1.4 cm. Other grains are moderately well sorted, angular and sub-spherical mono- and polycrystalline quartz (2.5 to 38.5 %), feldspar (0.5 %) and mica (<0.5 %). Micrite is locally present (0 to 3 % of samples). In siltstone and sandstone interbeds, micrite comprises 35.5 to 63 % of samples. Clay-grade, orange-red coloured material (<1 %) has accumulated along siliciclastic grain boundaries (**Figure 5.7d**). Clay-grade material constitutes less than 1 % of the matrix of this facies.

A moderately diverse assemblage of fragmented and abraded bioclasts is present in the matrix. Upper surfaces of cobble-grade clasts may have oyster encrustations and cylindrical borings up to 4 mm in diameter (**Figures 5.5b** and **5.7b**). Fragmented and abraded bioclasts include bivalves (3.5 to 10.5 %), coralline algae (0 to 10 %), planktonic foraminifera (Globergerinids) (0 to 4 %), gastropods (1-2 %), echinoids (0 to 5 %) and serpulid tubes (< 5 %) (**Figure 5.7c** and **e**). *In situ* and overturned branching *Porites* colonies up to 80 cm in diameter occur on discrete bedding planes (**Figure 5.7a**).

Porosity and diagenesis

Diagenetic features of the calcareous lithoclastic conglomerate facies vary (cf. **Figures 5.7c** and **e**). Poorly developed micrite envelopes are observed on some carbonate intraclasts. Intergranular equant to drusy calcite spar with intragranular drusy and blocky calcite spar rim and pore-filling cements are common (**Figure 5.7e**). Echinoid spines commonly have syntaxial calcite overgrowth cements. Aragonitic bioclasts, including corals and some molluscan fragments, have been dissolved. The resultant mouldic macropores may have been occluded with secondary drusy calcite spar cement. Vuggy, mouldic and intergranular porosity comprises up to 38 % of this facies.

Interpretation: Depositional environment

Deposition of siliciclastics proximal to the sediment source is inferred from the textural and compositional immaturity of this facies. Lithic grains in the Fortuna Basin are sourced from the uplifted and eroded Internal Zone massifs (**see Chapter 4**) (Santisteban and Taberner 1988, Lonergan and Schreiber 1993, Gracés et al. 2001). Siliciclastic grains of the Altorreal section are likely to be sourced from the Carrascoy

Massif, based on proximity to the mountain range and the general northward progradation of conglomerates. Low-grade metasediments (quartz-mica schists, quartzites etc) and laminated limestones are characteristic of the Malaguide Complex that is exposed in the Carrascoy Massif. Minor erosion of relatively recently deposited sediments is inferred from the presence of intraclasts (marl, siliciclastic packstone and coral clasts) that are found in neighbouring areas of the Fortuna Basin (Santisteban 1981, Santisteban and Taberner 1988).

Rapid deposition of clasts from cohesive mass flows is interpreted from the very poorly sorted nature of beds (Nemec and Steel 1984, Mather 1993). Coarsening upward trends within beds is attributed to deposition from relatively poorly cohesive surging mass flow events (Nemec and Steel 1984, Mather 1993). Mather (1993) attributes the presence of silt and sandstone interbeds as evidence of the development of bipartite flows. Fluidization may be expected from rapidly deposited sediments, although it is noted by many authors that it is very difficult to fluidize very coarse-grained deposits as the intergranular porosity would allow easy escape of fluids (Lowe 1975, Postma 1983). Each individual bed is interpreted to correspond to a discrete flow event. This allowed colonisation by oysters (encrusting boulders) and branching *Porites* colonies during quiescent periods.

Mass flow deposits are characteristic of fan delta front environments (Postma 1984, Mather 1993). The steep cross-stratification characteristic of this facies is interpreted to represent the prograding palaeoshelf profile. Haematite coatings on siliciclastic grains, in addition to the presence of carbonate nodules (caliche), has been interpreted by Mather (1993) to indicate pedogenic alteration. Caliche are absent in the Altorreal study area, thus it is suggested that either 1) there was insufficient time to allow pedogenic processes beyond haematite-coating of grains or 2) another process is responsible for the coatings.

It is concluded that this depositional system prograded into a marine body of water from the presence of marine fauna (bivalves, echinoids and planktonic foraminifera). High-energy conditions are confirmed from the moderate abundance of robust, encrusting oysters and fragmented and abraded bioclasts in the matrix.

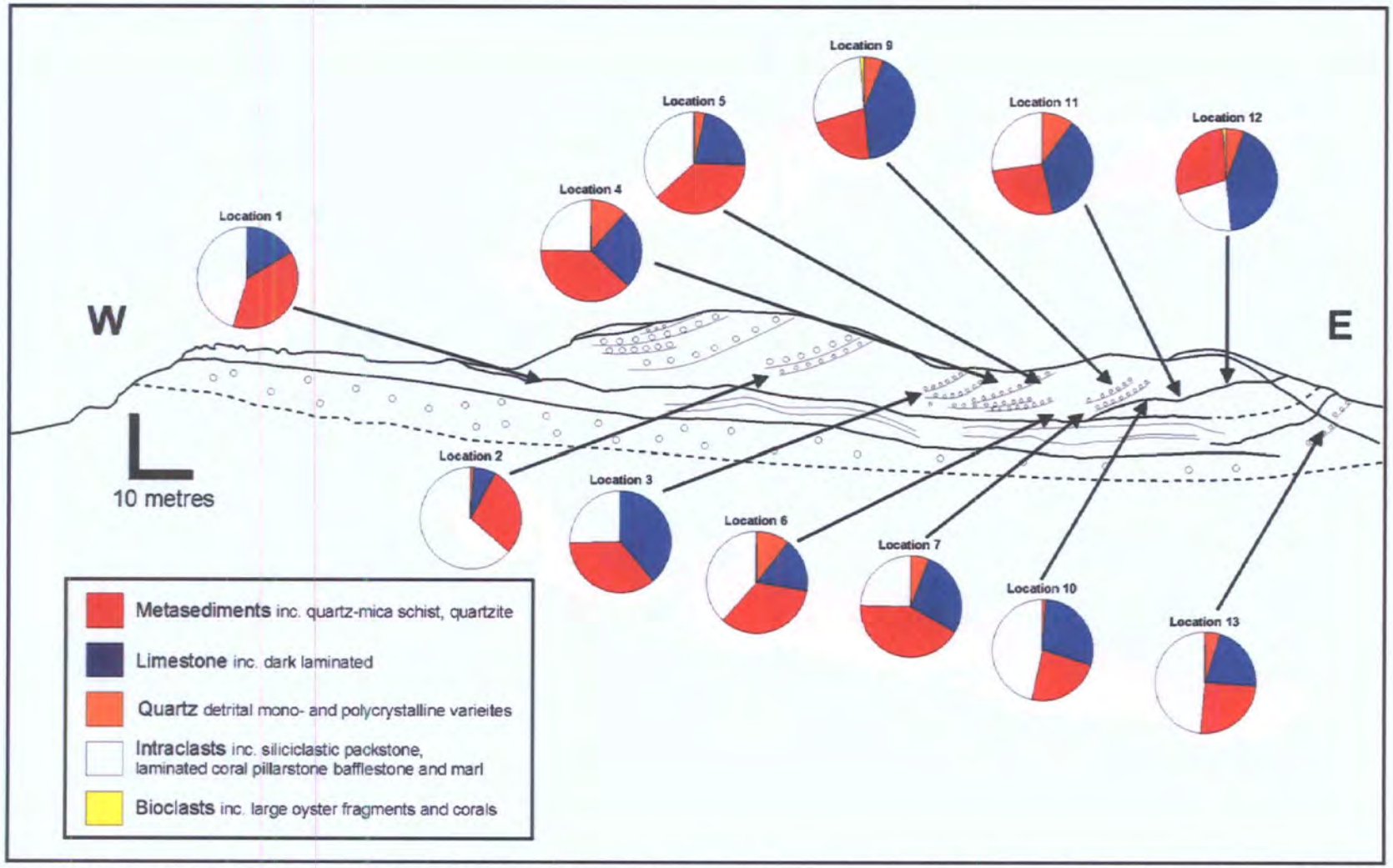
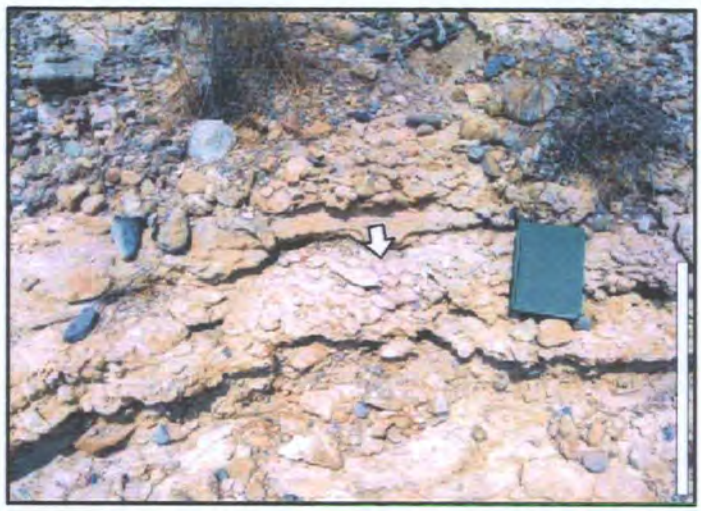
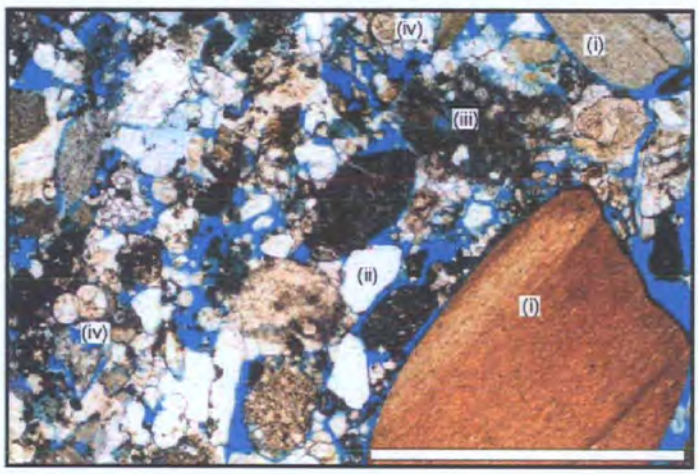


Figure 5.6 Figure to illustrate the composition of pebble to boulder-grade clasts along the exposed profile of the calcareous lithoclastic conglomerate facies. No significant trends are identified.





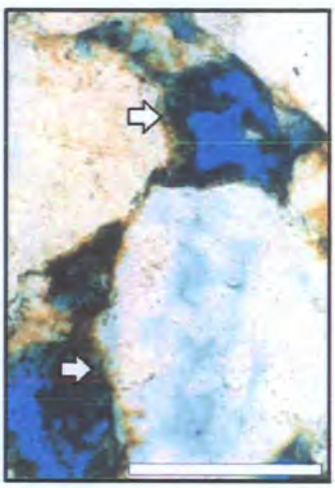


Figure 5.7a Isolated *in situ Porites* colony situated within the calcareous lithoclastic conglomerate facies (log FA-9, bed 10). Scale bar=50 cm.

Figure 5.7b Encrustation of a cobble-grade metasediment clast by robust oysters (log CA-9, bed 10). Pen=12cm

Figure 5.7c, d and **e** Photomicrographs of the calcareous lithoclastic conglomerate facies. **c)** (PPL) A high-porosity sample (porosity in blue) containing abundant lithics (i) and quartz (ii) grains, with patchy micrite (iii) and planktonic foraminifera (iv). Scalebar=2mm (sample LFA 66, log FA-11 bed 11). **d)** (PPL) Detailed view of haematite-coatings around siliciclastic grains (sample LCA 64, log FA-9, bed 10). Scalebar=0.5 mm. **e)** (XPL) Calcite cemented sample containing abundant oyster fragments (i) and serpulids (ii). Calcite cement (iii) is coarse with blocky equant crystals (sample LFA 28, log FA-5 bed 5). Scalebar=2mm.

Proximity to an area of coral development is inferred from the presence of large coral-microbial blocks that are found adjacent in some outcrops. In the Sorbas Basin, Braga and Martín (1996) infer the destruction of previous reef deposits (by siliciclastic debris flows) from the presence of reef blocks and reef-derived debris within fan delta conglomerates.

In summary, the calcareous lithoclastic conglomerate facies represents moderate to very high energy, intermittent deposition of siliciclastic material in a proximal marine fan-delta environment. Although the actual areal extent of the fan was not calculated in this study, comparable fan bodies of the Sorbas Basin are hundreds of metres across (Braga and Martín 1996). This facies is comparable to the Fan Delta facies of the eastern Fortuna Basin, adjacent to the Sierra Espuña described by Lonergan and Schreiber (1993) and Mankiewicz (1995). In addition, similar proximal fan delta facies are documented from the Miocene of the Sorbas Basin (Braga and Martín 1996), the San Miguel de Salinas Basin (Reinhold 1995), the Granada Basin (Martín *et al.* 1989, Braga *et al.* 1990) the Guadix Basin (Soria *et al.* 1999), the Lorca Basin (Wrobel and Michalzik 1999) and the Llucmajor Area of Mallorca (Pomar 1991, Pomar *et al.* 1996). In the Guadix and Lorca Basins, large prograding clinoforms and lense-shaped bed morphologies of bioclastic conglomerates are interpreted as the foresets and bottomsets of Gilbert-type deltaic systems that were located along basin margins (Soria *et al.* 1999 and Wrobel and Michalzik 1999). A distributary pattern with steep gradients and channels with a high bedload ratio is inferred from the coarse-grained nature of siliciclastics in fan deltas of the Lorca Basin (Wrobel and Michalzik 1999). Soria *et al.* (1999) interpret deposition of the bioclastic conglomerates of the Guadix Basin in extremely shallow marine conditions (close to fair weather wave base) during low sea level. Subaerial emergence is interpreted from the presence of palaeosols at the top of fan delta conglomerate sequences in the Granada Basin by Braga *et al.* (1990), although this is not seen in the Altorreal area.

5.2.1.2 Calcareous litharenite

Lithologies: Calcareous litharenite
Pebbly litharenite

Occurrence and bed characteristics

This facies is encountered in most logged sections in the Altorreal area. This facies often occurs below and intercalated with the calcareous lithoclastic conglomerate facies and above the siliciclastic bioclastic packstone facies (**Figures 5.4b and 5.8a**). Lower contacts of the calcareous litharenite facies with the siliciclastic bioclastic packstone facies are erosional (**Figure 5.4b**). Upper contacts with the calcareous lithoclastic conglomerate are erosional (**Figure 58.b**).

Beds occurs as tabular (**Figure 5.4b**) and lenticular units (**Figure 5.8a**). Tabular beds are up to 2 m thick and extend laterally in excess of 100 m. Tabular beds demonstrate relative shallow inclination compared to conglomeratic units. Poorly developed symmetrical ripple lamination is observed on discrete bedding surfaces, although this facies is largely massive. Ripple height is up to 3 cm, and ripple crests are sharp, sinuous and bifurcating. Small lenticular beds are up to 0.3 m thick and 2 m long (**Figure 5.8a**). Trough cross-stratification is observed in smaller sandstone bodies (**Figure 5.8a**).

The calcareous litharenite facies is locally pebbly and beds may be characterised by rapid fining upward with localised clast imbrication. In addition, randomly orientated, fragmented and abraded branching coral fragments (probably *Porites*) up to approximately 10 cm in length may occur (**Fig 5.8b**). 'Crinkly' laminae are observed on the upper surfaces of beds of this facies (**Figure 5.8b**). Laminae are defined by concentration of silt-grade grains.

Lithological description

Exposed surfaces of the calcareous litharenite facies weather to a pale brown to buff colour (**Figures 5.8a** and **b**). Fresh surfaces are a buff colour. This facies comprises subangular to well-rounded quartz (17.5 to 30 %), metasediments (1.5 to 15 %), limestone (0 to 7 %) and calcite cement fragments (< 3 %) (**Figures 5.9a and b**). Micrite matrix is locally preserved, comprising 3 to 47.5 % of samples. Non-carbonate clay-grade material rich in haematite often coats grains and may be concentrated along irregular laminae (**Figure 5.9b**).

Figure 5.8a Lenticular calcareous litharenite unit (arrowed). Log FA-6, bed 8). Scalebar=50 cm.

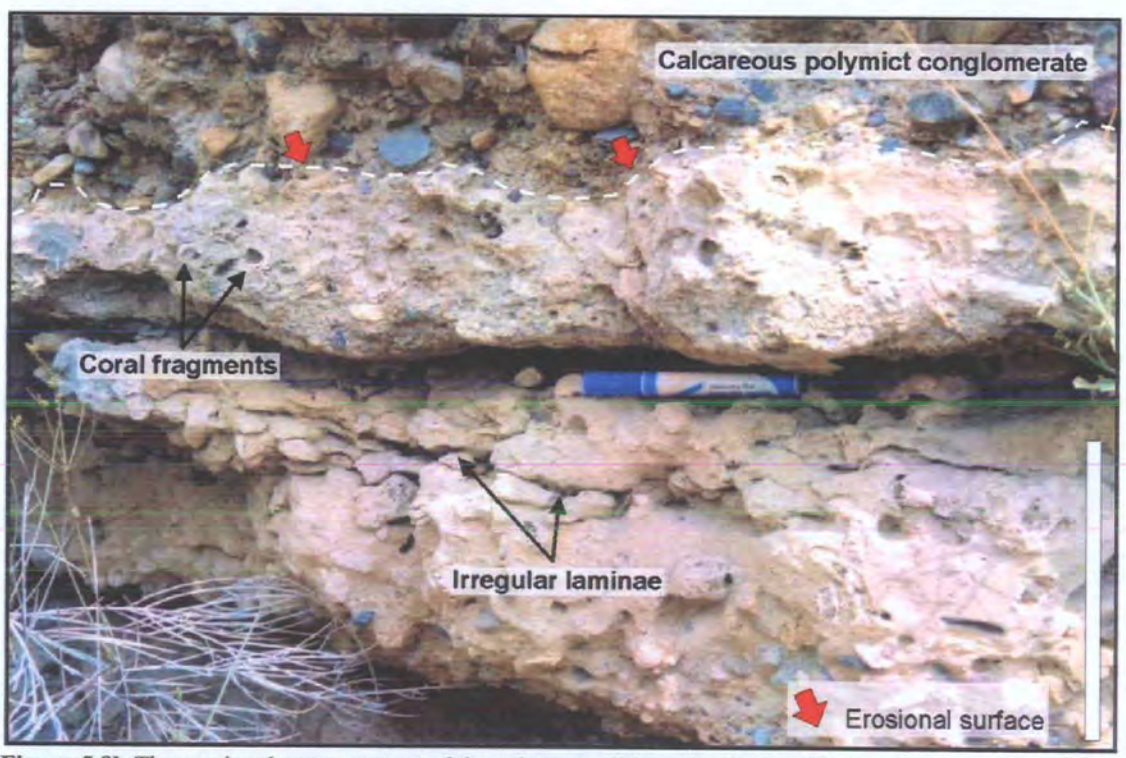


Figure 5.8b The erosional upper contact of the calcareous litharenite facies with the calcareous lithoclastic conglomerate (arrowed). The calcareous sandstone facies at this locality contains abundant reworked *Porites* branching coral fragments (indicated). Irregular laminae are also evident (log FA-11, bed 4). Scalebar=50 cm.

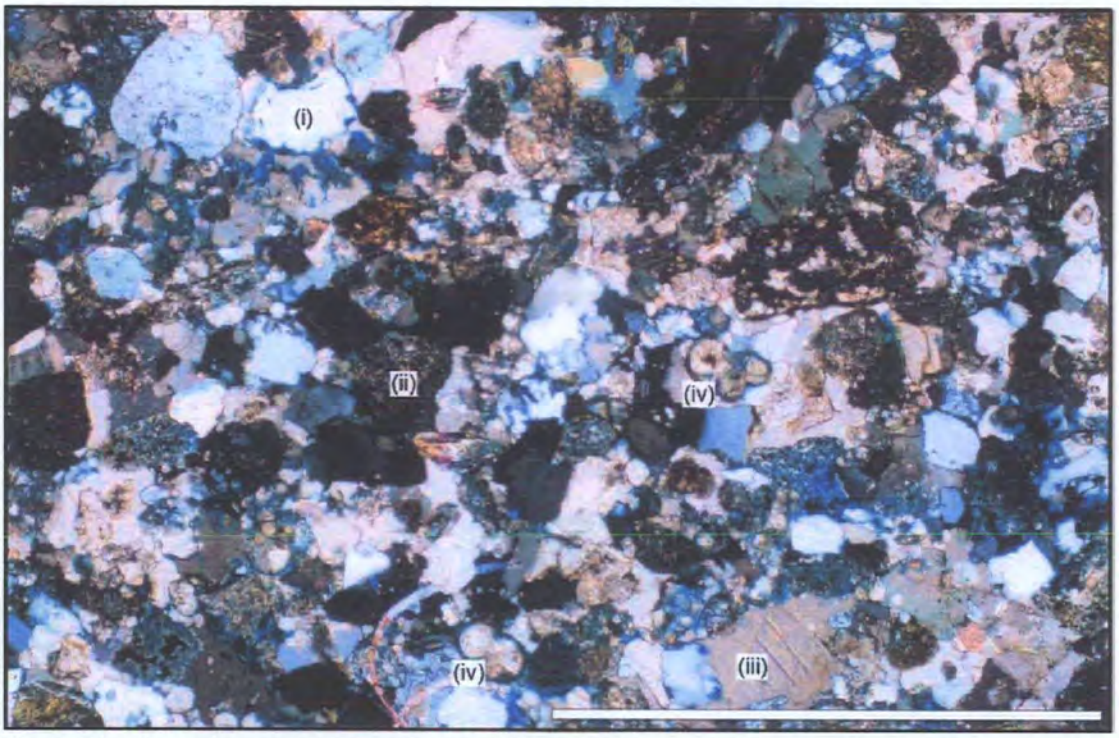


Figure 5.9a Photomicrograph of the calcareous litharenite facies. This facies contains angular, moderately well-sorted quartz (i) and lithic grains (ii). Calcite grains (iii) demonstrate syntaxial overgrowth cements. Planktonic foraminifera (iv) are also present (sample LFA 70, log FA-8, bed 11). Scalebar=50 cm.

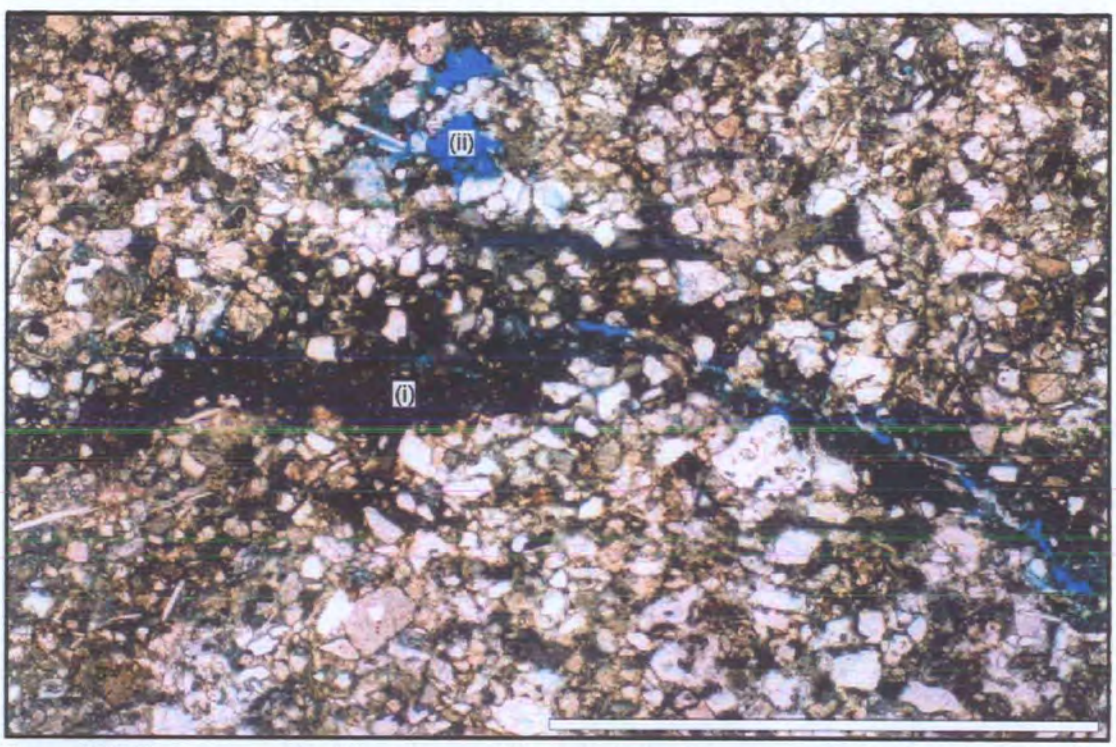


Figure 5.9b Photomicrograph of the calcareous litharenite facies. In many samples, grains are coated with a haematite-rich clay that is yellowish in thin section. Clays are also concentrated along irregular laminae (i). Primary intergranular and secondary vuggy pores (ii) are preserved (sample LFA 65, log FA-9, bed 9). Scalebar=50 cm.

The calcareous litharenite facies contains a moderately diverse fossil assemblage comprising bivalves (2 to 10 %), gastropods (0.5 to 1 %), echinoids (1.5 to 5 %), planktonic foraminifera (0.5 to 3.5 %), *Amphistegina* (<0.5 %), coralline algae (<0.5 %) and small hyaline benthic foraminifera (< 0.5 %) (**Figure 5.9a**). Bivalves are dominated by robust oysters, with pectens more abundant in finer-grained sediments. *Amphistegina* tests up to 0.5 mm in diameter have a robust, almost spherical morphology. Most fossils are fragmented and abraded.

Porosity and diagenesis

Originally aragonitic bioclasts (corals and gastropods) have been leached and the resultant mouldic pores have been locally occluded with blocky calcite cement. Many biomouldic pores have remained open. In comparison, oysters are well preserved. Calcite overgrowth cements (0.5 to 1.5 %) have developed on calcite grains and blocky calcite cements have developed within intergranular pores (**Figure 5.9a**). Up to 20 % of sediments comprise primary intergranular and secondary mouldic porosity.

Interpretation: environment of deposition

The composition of sediments of the calcareous litharenite facies is comparable to that of the calcareous lithoclastic conglomerate facies (**Section 5.2.1.1**), thus sediments are derived from the same source. However, deposition within a setting characterised by relatively low gradients is inferred from the low inclination of beds. Transport and reworking of coarse sand to pebble-grade siliciclastics prior to deposition and lithification is inferred from the well-rounded nature of grains. Tabular beds are largely massive, although the presence of wave-ripple lamination indicates deposition above fair weather wave base (Reading 1996).

Normal marine conditions with open circulation are inferred from the presence of planktonic foraminifera and the modest diversity of the bioclast assemblage. Most bioclasts are fragmented and abraded, thus have undergone transport and reworking prior to deposition and lithification. The robust, almost spherical forms of *Amphistegina* are characteristic of high-energy, shallow water conditions (Leutenegger 1984, Hallock and Glenn 1985) thus an input of erosional detritus from a high-energy shallow marine environment is inferred. Broadly low to moderate

energy conditions are inferred from the micrite matrix, although reworking by marine currents at depths above fair weather wave base is suggested from wave-ripple laminae. It is concluded that tabular calcareous litharenite beds that occur stratigraphically below the calcareous lithoclast conglomerate facies represent the relatively distal low-energy equivalent of the conglomerates.

Trough cross-stratification, present within smaller lenticular sand bodies is characteristic of alluvial sedimentary settings, and represents migration of three-dimensional dunes (Reading 1996). The lenticular calcareous litharenite units situated within the calcareous lithoclastic conglomerate facies (**Figure 5.9b**) are interpreted to represent small (dm-scale) channel systems. Similar features described by Mather (1993) are interpreted as in-filled scours.

It is suggested that the crinkly laminar features observed on some beds of the calcareous litharenite facies (**Figure 5.8b**) are microbial in origin (i.e. a microbial mat) although no primary microbial structures remain. Microbial communities may produce mats with crinkled morphologies as well as flat mats (Logan *et al.* 1974, Kinsman and Park 1976). Modern microbial development is particularly common in inter-tidal to shallow sub-tidal environments and the crinkly nature of laminae could also be a consequence of subaerial exposure and dessication of a microbial mat (Riding 1991, 2000, Tucker and Wright 1990). Haematite-rich coatings on grains have been used to infer subaerial exposure and incipient soil development (Mather 1993). Nofke (1998) suggests that the growth of microbial mats occurs during pauses in sedimentation. Microbially laminated sandstones are not described from any of the other Neogene basins of SE Spain or the Mediterranean. Microbialites have been described from modern lithified lacustrine alluvial fan sediments of Nevada (Osborne *et al.* 1982).

In summary, the calcareous litharenite facies formed within a low to moderate-energy fan-delta front to prodelta setting, and is interpreted to be the relatively distal equivalent of the calcareous lithoclastic conglomerate facies. Bioclast-rich calcareous sandstones closely associated with conglomeratic facies are described from the Granada Basin (Braga *et al.* 1990), Guadix Basin (Soria *et al.* 1999), Lorca Basin (Wrobel and Michalzik 1999) and the Sorbas Basin (Braga and Martín 1996). Bioclastic sands of the Lorca Basin are interpreted as delta front to prodelta mouth bar sands, coastal barrier sands and subaqueous levées (Wrobel and Michalzik 1999).

5.2.2 Mixed carbonate-siliciclastic facies

These facies are characterised by a mixed carbonate siliciclastic composition. The textures present in this grouping are clastic-rich marl and packstone. Facies present in this grouping are characterised by a low/negligible siliciclastic influx.

5.2.2.1 Silty marl

Lithologies: Silty marl

Occurrence and bed characteristics

The silty marl facies is present at the base of all measured sections. The thickness and nature of the lower contact of this facies cannot be determined, although Lonergan and Schreiber (1993) and Mankiewicz (1995) describe an erosional lower contact with Betic Internal Zone rocks in the western parts of the Fortuna Basin. The silty marl facies is always succeeded by the calcareous lithoclastic conglomerate facies in the Altorreal ridge sections, and is not observed in association with any of the other facies present in the study area (**Figure 5.2**). The upper contact with the calcareous lithoclastic conglomerate facies is undulating and erosive (**Figure 5.4a**). The silty marl facies is largely massive with localised mm-scale laminations.

Lithological description

Exposed surfaces of this facies weather to a white friable mud (**Figure 5.10a**). Fresh surfaces are a pale grey colour. Up to 48 % of the silty marl facies comprises homogeneous micrite, with minor silt-grade quartz (<3.5 %) and quartz-mica schist (<1 %) (**Figure 5.10b**). Grains are moderately well sorted, angular with low sphericity. Grains are predominantly detrital quartz with minor quartz-mica schist and calcite (**Figure 5.10b**). This facies contains a modest biota dominated by planktonic foraminifera (< 3.5 %) and echinoids (< 1%). Planktonic foraminifera tests are preserved intact and range in size from 0.01 to 0.05 mm (**Figure 5.10c**).

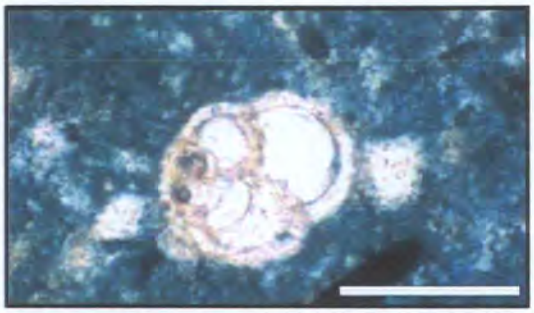


Figure 5.10c Detailed view of the silty marl facies. This facies is characterised by the presence of planktonic foraminifera (sample LFA 66-m, log FA-11, bed 1). Chambers have been cemented with poikilotopic calcite (clear in thin section) Scalebar=0.25 mm

Figure 5.10a Field exposure of the silty marl facies (GR62201385, facing E). This facies weathers to a white friable mud. Sediments are often crosscut by gypsiferous veins (arrowed), interpreted as a late diagenetic feature.

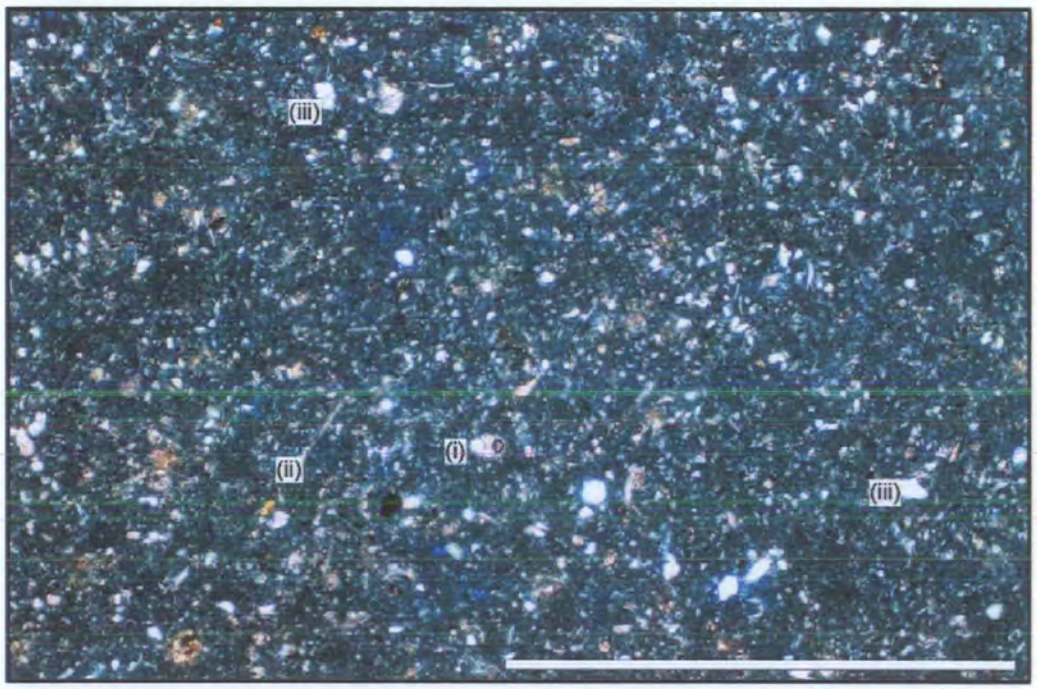


Figure 5.10b Thin section photomicrograph of the silty marl facies. This facies is characterised by planktonic foraminifera (i) and rare echinoids (ii) situated within microporous micrite. Microporosity is indicated from the blue hue of the thin section. Siliciclastic grains (iii) are silt-grade, angular and very well sorted (sample LFA 66-m, log FA-11, bed 1). Scalebar=2 mm.

Diagenesis

Gypsum-vein networks can be observed in some localities (**Figure 5.10a**). Veins crosscut diffuse laminations and may extend laterally up to 2 m. The silty marl facies contains up to 20 % matrix microporosity (**Figure 5.10b**). Poikilotopic calcite cements are observed within foraminifera tests (**Figure 5.10c**).

Interpretation: environment of deposition

Open marine depositional conditions are inferred for this facies from the presence of planktonic foraminifera and rare echinoid spines. Constraining the depth of this facies is difficult because of the apparent absence of sedimentary structures and ichnofabrics although low-energy conditions and deposition at/below storm wave base are inferred from the abundance of fine-grained micrite and localised presence of diffuse, fissile laminations. It is interpreted that fine-grained siliciclastic material was deposited from suspension. Limited abrasion prior to deposition and lithification is inferred from the angular nature of siliciclastic grains.

In summary it is interpreted that this facies was deposited in an open marine, low-energy distal environment, below storm wave base, with a moderate siliciclastic influx. This facies has affinities with Standard Microfacies 3 of Wilson (1975). Comparable marls rich in planktonic foraminifera are described from the Guadix Basin (Soria *et al.* 1999), Sorbas Basin (Braga and Martín 1996) and the Alicante-Elche Basin (Calvet *et al.* 1996). In addition, turbidites and distinct slump horizons are described within the marl horizons of the Guadix, Alicante-Elche and San Miguel de Salinas Basins (Soria *et al.* 1999, Calvet *et al.* 1996 and Rheinhold 1997). This is not observed in the marl of the Altorreal study area, and a relatively flat depositional environment is inferred.

5.2.2.2 Siliciclastic bioclastic packstone

Lithologies: Siliciclastic bioclastic packstone
Molluscan packstone
Bioclastic packstone

Occurrence and bed characteristics

The siliciclastic bioclastic packstone facies is encountered in the middle part of the studied succession (**Figure 5.2**). This facies overlies the calcareous lithoclastic conglomerate facies, and may be intercalated with the calcareous litharenite facies. This facies is also observed to pass laterally into the laminated coral platestone facies (**Figure 5.2**). Measured bed thickness ranges from 0.4 to 3.3 m with typical thicknesses around 1.4 m. Beds of this facies are typically horizontally stratified (**Figure 5.2**). Upper and lower bed contacts are typically poorly defined, and beds appear to amalgamate along section yielding total thickness of this facies up to 4.4 m (e.g. log FA-11, bed 3). Lower bedding contacts with conglomerates are gradational (**Figure 5.11a**). Poorly sorted, sub-angular and sub-spherical cobble-grade clasts up to 50 cm may occur at the base of this facies (i.e. FA-11, bed 5, **Figure 5.11a**). Beds extend laterally along the entire east-west length of the main ridge (> 200 m).

Lithophagid borings up to 0.5 cm in diameter (**Figure 5.11b**) are common on the upper surfaces of a bed of this facies that precedes the calcareous lithoclastic conglomerate facies (e.g. log FA-10 bed 10, log FA-11 bed 3) (**Figure 5.4b**). *In situ*, warty-branching coralline algae occur locally on the bored horizons (**Figure 5.11b**).

Lithological description

Exposed surfaces of this facies weather to a pale brown-buff colour. Fresh surfaces are a pale buff colour. Up to 10 % of this facies comprises coarse sand to pebble grade siliciclastic grains. Pebble-grade grains up to 1.5 cm are poorly sorted, well rounded and sub-spherical.

The siliciclastic bioclastic packstone facies contains a moderately diverse fossil assemblage, comprising oysters (7 to 25 %), gastropods (0 to 3 %) coralline algae, bryozoa (0 to 2 %), coral fragments (< 0.5 %), worm tubes (< 0.5 %), planktonic foraminifera (0.5 %), miliolids (0 to 1 %), *Amphistegina* (<0.5 %), victoriellids (0 to 1 %) and undifferentiated smaller benthic foraminifera (< 0.5 %)

Figure 5.11a Outcrop of the siliciclastic bioclastic packstone facies (log FA-11, bed 5). Where this facies succeeds the calcareous lithoclastic conglomerate facies, outsized boulders (in this instance a limestone clast-arrowed) are present at the base. Lithic boulders fine upwards over 50 cm. Scalebar=50 cm.

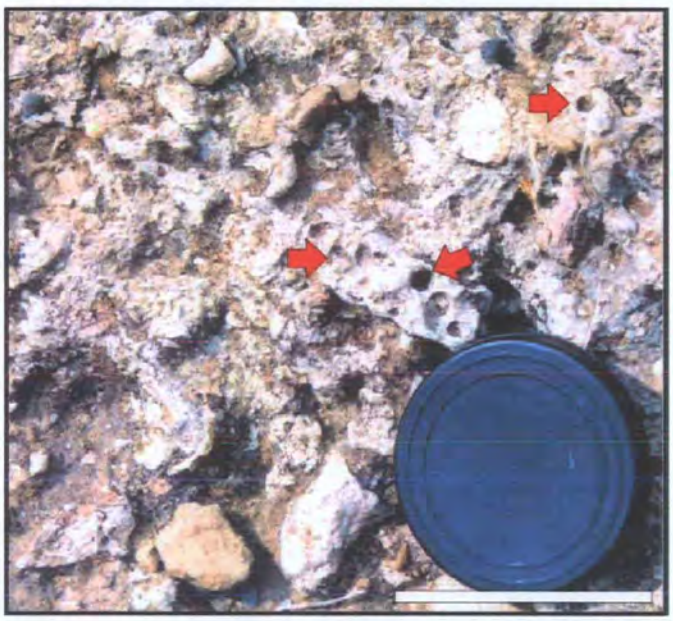


Figure 5.11b Outcrop of the siliciclastic bioclastic packstone facies (log FA-10, bed 10). It is interpreted that abundant lithophagid borings on the upper surface of this bed (arrowed) represent a hardground. See text for details. Scalebar=5 cm.

Figure 5.11c Fragmented and abraded oysters within the siliciclastic bioclastic packstone facies (isolated exposure within the Casa de Penalver area-location indicated on **Figure 5.1**). Scalebar=5 cm.

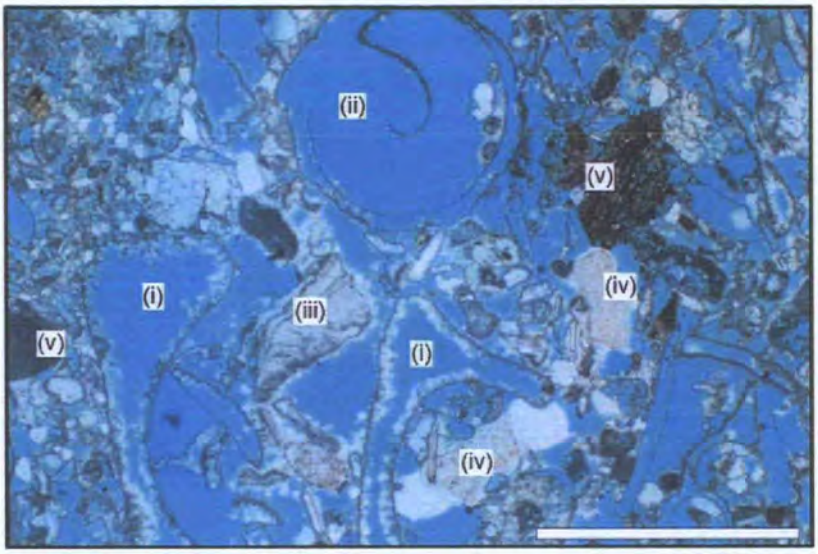


Figure 5.12a Photomicrograph (PPL) of the siliciclastic bioclastic packstone facies (sample LFA 32, log FA-6, bed 3). Most bioclasts in this facies have been leached, although moulds of bivalves (i) and gastropods (ii) are recognized. Other clasts include well-preserved oysters (iii), echinoids with syntaxial overgrowths (iv) and lithic grains (v). Scalebar=2mm.

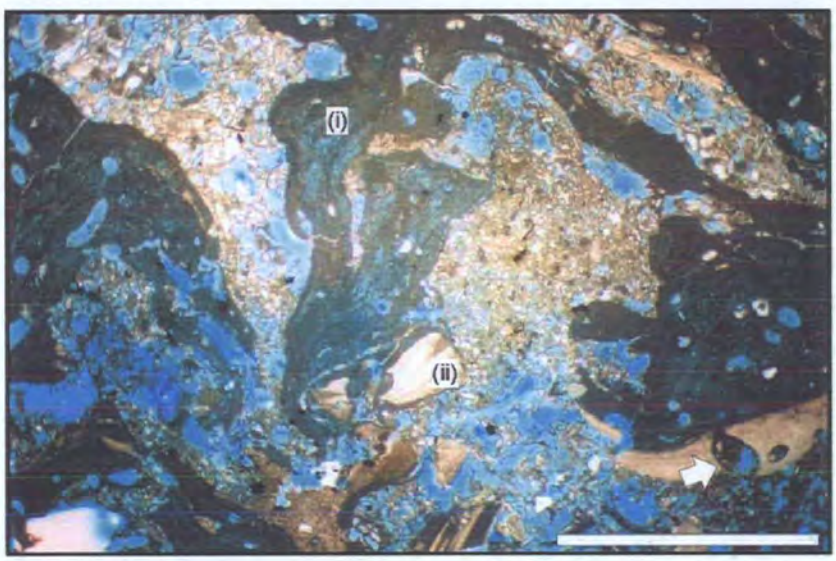


Figure 5.12b Photomicrograph (PPL) of the siliciclastic bioclastic packstone facies (sample LFA 33, log FA-6, bed 5). Laminar encrusting coralline algae (i) with algae-encrusted oyster fragments (ii). Scalebar=2mm.

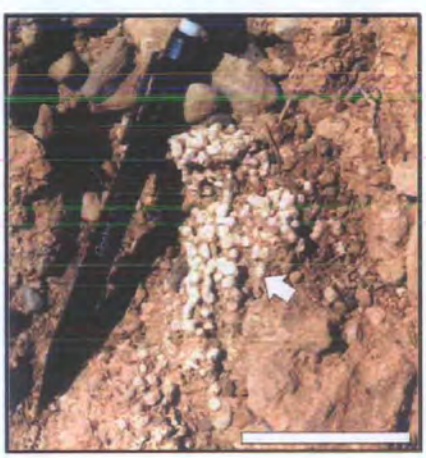


Figure 5.12c *In situ* warty branching coralline algae situated on the upper surface of the siliciclastic bioclastic packstone facies (log FA-8, bed 10) Scalebar=2mm.

(**Figure 5.12a**). Planktonic foraminifera up to 0.25 mm are preserved intact. Miliolids up to 1.5 mm are typically preserved intact, although tests are highly abraded with a dusky appearance. Robust, intermediate *Amphistegina* tests up to 2 mm are relatively rare (less than 1 % bioclasts) and may be fragmented and abraded. Victoriellids are very rare (less than 1 % of bioclasts). Most fossils are fragmented and partially abraded. Oyster fragments are up to 2.5 cm across (**Figure 5.11c**). Coralline algae are present as elliptical, concentric laminar rhodoliths up to 2 cm in diameter. Rhodolith nuclei are bivalve and bryozoan fragments. Coralline algae also occur as fragmented, laminar crusts that are detached from their substrate, and *in situ* warty growths (**Figure 5.12b** and **c**).

Micrite is rare in most samples of this facies, although it is locally abundant comprising up to 50 % of a sample. Siliciclastic grains are mono- and polycrystalline quartz (1 to 22 %), metasediments (2 to 20.5 %) and laminated limestone (2 to 30 %).

Peloidal, silty interbeds are composed predominantly of dark brown, peloidal micrite (63.5 %) with quartz (5.5 %), metasediments (12.5 %), laminated limestone clasts (3.0 %) also containing abraded coralline algae (5.5 %), bivalves (2.5 %), echinoids (1.5 %) and planktonic foraminifera (0.5 %) (**Figures 5.13a** and **b**).

Diagenesis

Originally aragonitic organisms have been leached, and resultant mouldic pores have been partially occluded with blocky calcite rim cement (10 to 22.5 %) (**Figure 5.12a**). Up to 55 % of the siliciclastic bioclastic packstone facies is secondary mouldic porosity with locally preserved intergranular primary porosity (**Figures 5.12 a** and **b**). It is postulated that significant leaching of the matrix has also occurred. Leaching of sediments is particularly prevalent along the *Lithophaga*-bored horizon (**Figure 5.2**).

Interpretation: environment of deposition

A marine depositional environment with open marine circulation is interpreted from the presence of planktonic foraminifera. Deposition within, or close to, the photic zone is inferred from the presence of coralline algae and *Amphistegina* (Ghose 1977, Hallock and Glenn 1986). Laterally extensive coralline algae are

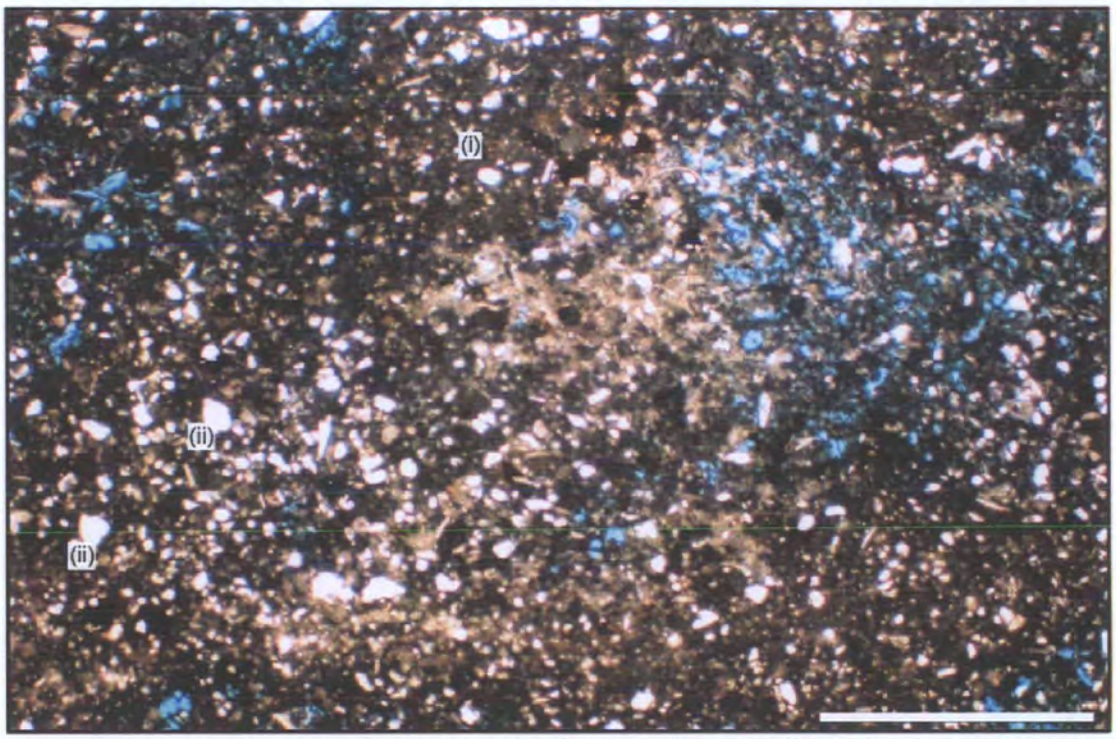


Figure 5.13a Photomicrograph of a peloidal, silty interbed within the siliciclastic bioclastic packstone facies. This sediment is dark in ppl, with dark brown peloidal micrite (i) with silt-grade, angular well-sorted siliciclastic grains (ii) (sample LFA 44a, log 8, bed 8). Scalebar=2 mm.

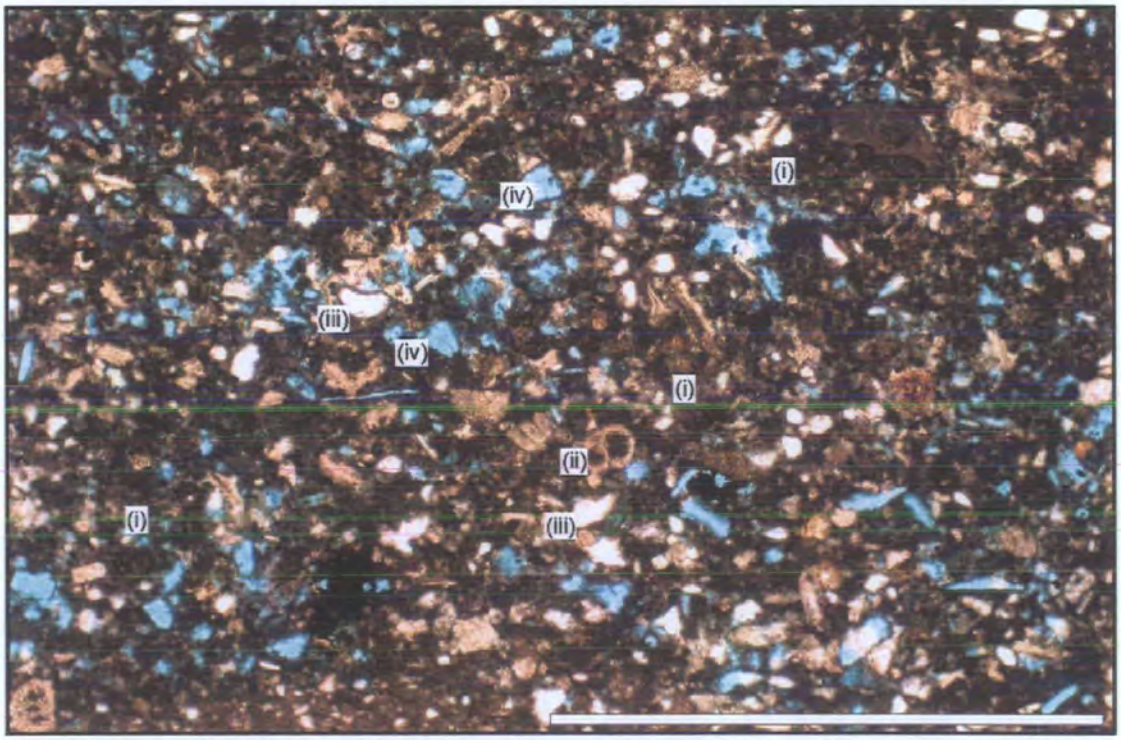


Figure 5.13b Detailed photomicrograph of the above sample (sample LFA 44a, log 8, bed 8). Sediment is dominated by peoidal micrite (i) with planktonic foraminifera (ii) and angular siliciclastic grains (iii). Porosity is dominated by mouldic pores (iv), which are combination of leached calcite grains and bioclasts. Scalebar=2 mm.

consistent with moderate to lower-energy settings, where competition for space is limited (Minnery *et al.* 1985, Minnery 1990).

Proximity to a siliciclastic source and/or a noticeable siliciclastic input is inferred from the significant percentage of coarse, siliciclastic material within sediments. Transport and/or re-working of siliciclastics prior to deposition and lithification is inferred from the well-rounded morphology of pebble-grade lithics, although beds of this facies are largely massive, with little evidence of continuous wave or current re-working (i.e. ripple lamination). However, periodic winnowing by currents or variations in the depositional energy is inferred from the localised textural variability of this facies. In addition, most bioclasts are fragmented and abraded, indicating transport and reworking prior to deposition and lithification. It is inferred that robust oyster fragments are derived from a nearby high-energy environment comparable to the calcareous lithoclast conglomerate facies where oysters are observed *in situ* encrusting boulders (see **Section 5.2.2.1**).

A period of very slow deposition, syn-depositional cementation and subsequent boring is inferred from the presence of lithophagid borings on the top surface of the bed illustrated on **Figure 5.4b**. This feature is not observed on the upper surface of other beds of this facies. This surface is interpreted as a hardground. Recent hardgrounds form just below the sea floor, and typically develop in areas of slow sedimentation and high current activity (Tucker and Wright 1990). Boring of hardgrounds by lithophagid bivalves is common (Tucker and Wright 1990). Cementation of this facies and hardground development was probably aided by its high porosity that would ease movement of porewaters through sediment and periodically agitated energy conditions. It is inferred that blocky calcite rim cements (**Figure 5.12a**) developed soon after deposition within the marine diagenetic environment. These cements are particularly well developed in samples from the packstone horizon illustrated on **Figure 5.4b**.

In summary, the siliciclastic bioclastic packstone facies is interpreted as a moderate-energy, marine shelf deposit that accumulated within or very close to the photic zone in an area with a moderate siliciclastic and bioclastic influx with periodic reworking by currents. Comparable facies are described from the Granada Basin (Martin *et al.* 1989), the Guadix Basin (Soria *et al.* 1999) and the Lorca Basin (Wrobel and Michalzik 1999). Siliciclastic packstone beds in these basins have been

interpreted as the marine offshore-equivalents of coastal fan delta systems and as carbonates that developed on conglomeratic shoreline-attached platforms. Bioclastic carbonate facies of the Lorca Basin contain up to 15 % terrigenous siliciclastic material, and are interpreted as a shallow water carbonate-platform environment. However, although lithologically comparable, the siliciclastic packstones of the Granada Basin show no evidence of early lithification and hardground formation such as inferred at Altorreal.

5.2.3 Carbonate-dominated facies

Facies belonging to this grouping are characterised by platestone, pillarstone and mixstone textures. Carbonate facies occur as sigmoidal clinoforms that pinch out laterally perpendicular to dip direction units (**Figure 5.2**). Two carbonate intervals situated within siliciclastic sediments are identified on the northern and southern ridge sections (**Figures 5.2, 5.3** and **5.14**). A minor interval is identified in the Casa de Penalver area (**Figure 3.1**). Correlation of these is discussed in **Section 5.3**.

Measured lateral extent of the carbonate units perpendicular to the dip direction of the siliciclastic sediments is up to 50 m. Carbonate units extend laterally down-dip up to 100 m (**Figure 5.14**). Vertical thickness is up to 10 m. Bedding and other internal sedimentary structures are largely absent. There are distinct vertical and lateral transitional changes in coral colony morphology, with laminar, platy forms dominating at the base. These platy forms progressively change to predominantly head, massive and foliaceous forms in the middle and top of the carbonate units. Stick-like, branching forms occur towards the margins of carbonate bodies. Coral morphology can be used to determine relative water depths (Santisteban 1981, Santisteban and Taberner 1988, Esteban 1996). This zonation of colony morphology is unusually clear due to the low coral diversity typical of late Miocene Mediterranean reef systems (Esteban 1996). Zonation of colonial coral morphology can therefore be used (in conjunction with other sedimentological evidence) to reconstruct the environment in terms of relative water depth, siliciclastic influx etc. (Santisteban 1981, Grasso and Pedley 1988, 1989, Esteban 1996).

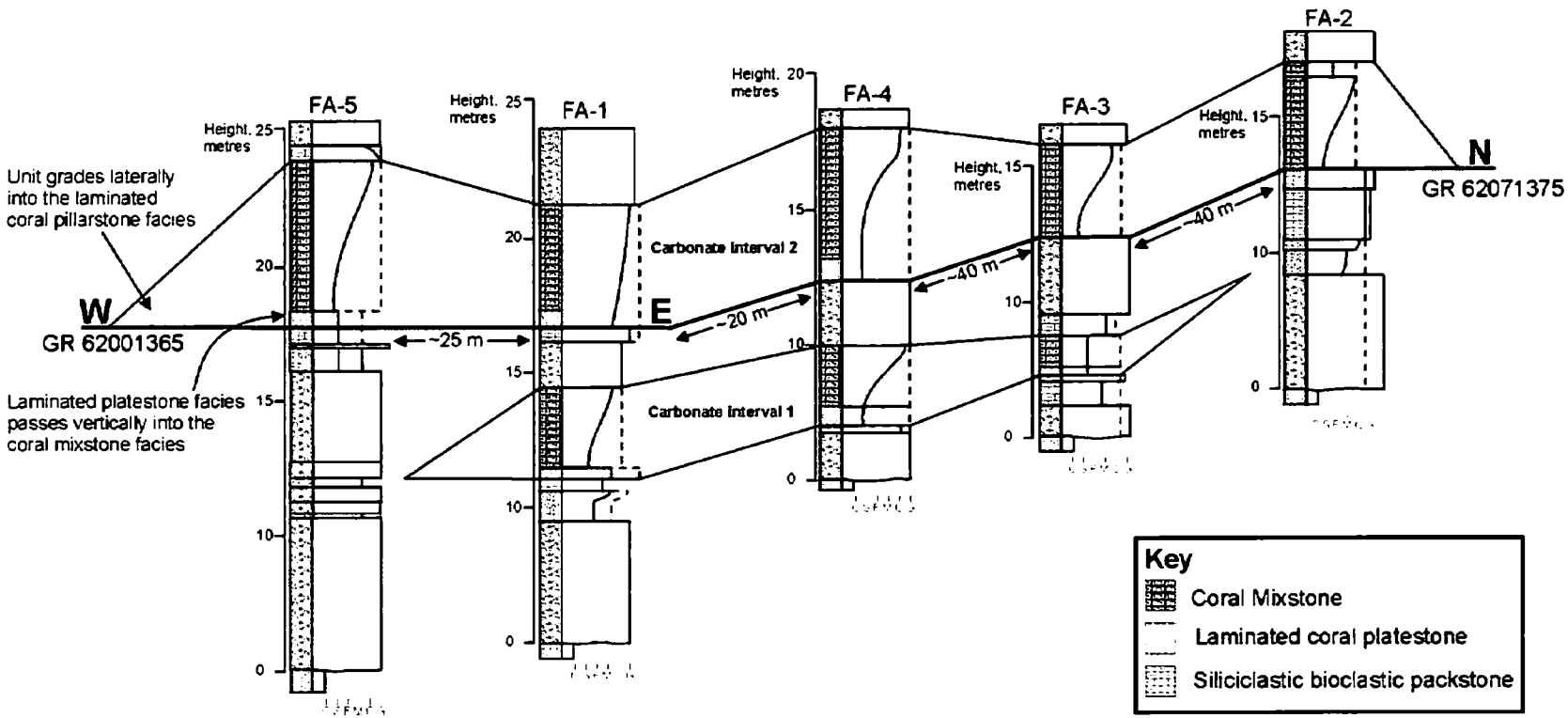


Figure 5.14 Correlation of selected logs from the northern ridge section at Altorreal illustrating the morphology and lateral continuity of carbonate units. Two isolated carbonate intervals (shaded grey for clarity) are identified in the northern ridge. The base of the carbonate intervals is typically marked by a thin (< 1m) interval of the laminated coral platestone facies that passes gradationally into the coral mixstone. Logs are "hung" from the base of the upper carbonate interval. All correlation tie lines have been traced in the field.

5.2.3.1 Laminated coral platestone

Lithologies: Laminated coral platestone

Occurrence and bed characteristics

This facies is not volumetrically important although it is encountered at the base of both carbonate intervals identified on the northern ridge section (logs FA-1, FA-2, FA-4 to FA-7). This facies is not encountered in the southern ridge section, or to the far south of the study areas at Casa de Peñalver (**Figure 5.1**). The laminated coral platestone facies typically comprises the lower metre of the carbonate unit and passes vertically into the laminated coral pillarstone and/or the coral mixstone facies. Upper contacts are gradational. The laminated coral platestone facies also passes laterally into the siliciclastic bioclastic packstone facies (**Figure 5.2**). The base of this facies may contain granule to boulder grade siliciclastic grains.

Lithological description

Exposed surfaces of the laminated coral platestone facies weather to a pale grey to white colour (**Figure 5.15a**). Fresh surfaces are white and have a finely crystalline texture. This facies is characterised by the presence of mm-scale, highly irregular sub-horizontal laminations that extend laterally for at least 1 m and in total are up to 75 cm thick (**Figure 5.15a**). In thin section, sub-mm scale dark brown micritic laminations alternate with orange-brown micrite with scattered siliciclastic grains (**Figure 5.15b**). Siliciclastics are well-sorted, sub-angular to well rounded and sub-spherical. Robust platy to massive corals are preserved towards the top of this facies (**Figure 515a**). Colonies are up to 75 cm across.

A bioclastic wacke/packstone matrix is locally preserved in this facies. The matrix comprises micrite (50.5 %), coral fragments (0.5 %), gastropods (5 %), echinoids (1.5 %), bivalves (8.5 %), bryozoa (0.5 %), coralline algae (1.5 %) and planktonic foraminifera (< 2 %). Siliciclastic grains are quartz (0.5 to 11.5 %), metasediments (<9.5 %) and limestone (2 %). Grains up to 0.25 mm are well sorted, angular and subspherical.

Figure 5.15a Outcrop of the laminated coral platestone facies (log FA-2, bed 7). Sediments are characterized by laterally extensive, irregular laminae (i) with sparse platy coral colonies (ii). Scalebar=50 cm.

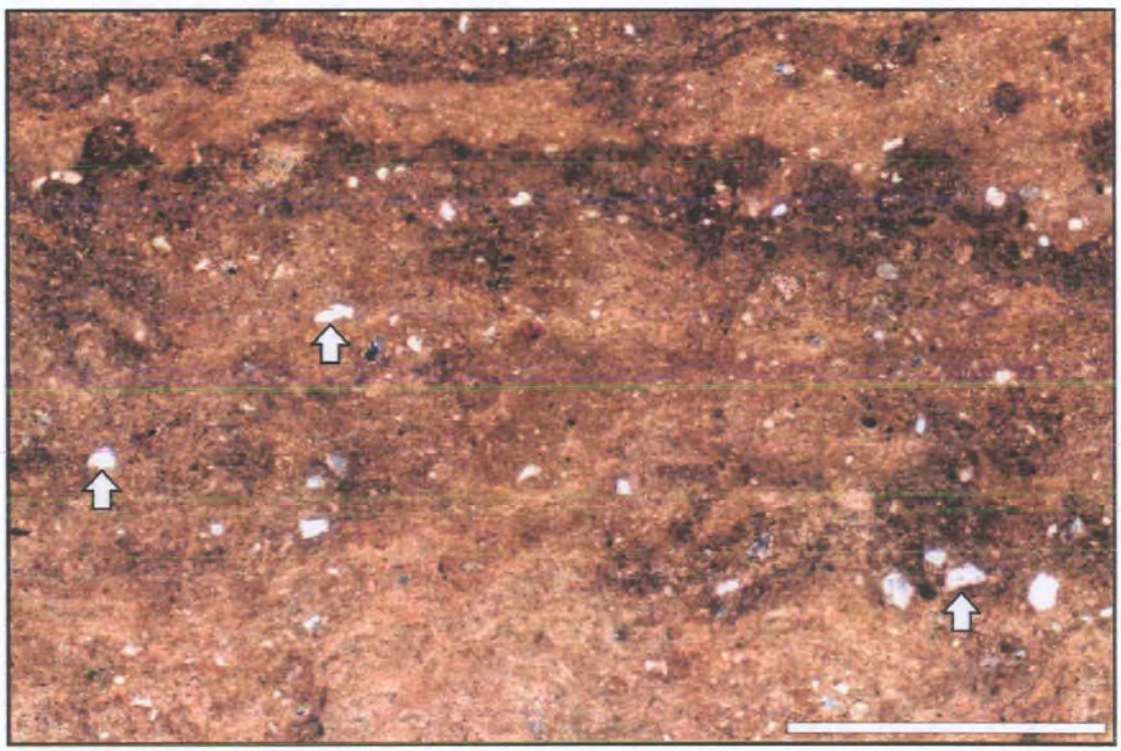


Figure 5.15b Detailed photomicrograph of the laminated coral platestone facies (sample LFA 16, log 2, bed 7). Sediment is characterized by irregular light and dark brown micritic laminae. Rare silt-grade siliciclastic grains (arrowed) occur randomly scattered throughout this sediment. Scalebar=2 mm.

Diagenesis

Intergranular drusy and equant calcite spar rim cement comprises up to 5 % of this facies. All aragonitic bioclasts (including corals and molluscs) have been dissolved and wholly, or partially, replaced with intragranular drusy calcite spar cement. Up to 30 % of the laminated coral platestone facies is vuggy, biomouldic and intergranular porosity.

Interpretation: environment of deposition

Irregular, mm-scale laminae defined by alternations of dark and light micrite are consistent with structures interpreted as microbial in origin by Esteban (1979), Riding (1991, 2002) and Nofke (1998). Using the classification scheme provided by Riding (2002), these structures are classified as stromatolites. No cellular microbial structures are preserved in thin section. The principal organisms involved in the growth of microbial biofilms and mats are bacteria (particularly cyanobacteria), small algae and fungi (Riding 2000). Noffke (1998) has suggested that growth of bacterial mats occurs mainly during pauses in sedimentation thus accretion of sediments on bacterial mats is likely to be episodic (Riding 2000). Modern microbial structures are ubiquitous on tidal flat areas (Tucker and Wright 1999). However, a subtidal environment of deposition is inferred for the laminated coral platestone facies since this facies includes *in situ* stenohaline biota. Steneck *et al.* (1998) suggest stromatolite development is restricted to environments where ecological processes such as colonisation, growth, competition and predation (herbivory) are low. Kendall and Skipwith (1968) attribute the abundance of microbial mats in the Persian Gulf lagoon to the limited grazing of invertebrates, perhaps as a consequence of reduced invertebrate populations due to enhanced salinities. Additionally Garrett (1970) has suggested that gastropod grazing was responsible for the apparent restriction of well-developed microbial mats in intertidal environments of the Bahama Banks. Reitner and Neuweiler (1995) suggest a more important factor promoting microbial development is nutrient-enrichment.

Normally, platy coral morphologies are consistent with environments with reduced light penetration, such as deep or turbid water (Santisteban and Taberner 1988, Pomar *et al.* 1996). However, coral colonies within this facies are thick suggesting agitated conditions. It is suggested from the association with laminar

microbialte that platy corals developed within very shallow water. It is postulated that horizontal growth predominated over vertical growth due to a lack in available accommodation space. Platy coral development is described from the shallow-water, high-energy reef crest zone of Miocene reefs in the Sorbas Basin (Riding 1991).

A moderate input of siliciclastics is inferred from the presence of siliciclastic grains in the matrix. Some transport and reworking of siliciclastic grains prior to deposition and lithification is suggested from the rounded nature of lithic grains in the matrix.

In summary, this facies represents the development of microbial mats and platy coral colonies within an agitated, very shallow-water subtidal setting. Similar facies are described from the Sorbas Basin (Riding 1991).

5.2.3.2 Coral mixstone

Lithologies: Coral mixstone
Coral mix/pillarstone

Occurrence and bed characteristics

The coral mixstone facies comprises the majority of the carbonate intervals identified in the northern ridge section (**Figure 5.14**). This facies is also encountered on the southern ridge section (**Figure 5.3**) and to the south at the Casa de Penalver section (**Figure 5.1**). The coral mixstone facies vertically succeeds the laminated coral platestone facies. The contact is gradational. The coral mixstone facies is also observed to overly the calcareous lithoclastic conglomerate and siliciclastic bioclastic packstone facies (**Figures 5.14 and 5.16a**). These contacts are sharp and undulating but non-erosive. The coral mixstone facies is typically succeeded by the calcareous lithoclastic conglomerate facies (**Figure 5.14**). The contact is erosional (**Figures 5.16b** and **c**). The maximum measured thickness of this facies is 10 m, although beds are observed to pinch out down-dip of underlying siliciclastic sediments (**Figure 5.14**).

Lithological description

Exposed surfaces of the coral mixstone facies weather to a pale brown-yellow colour (**Figures 5.16** and **5.17a**). Fresh surfaces are a pale grey-brown colour. This

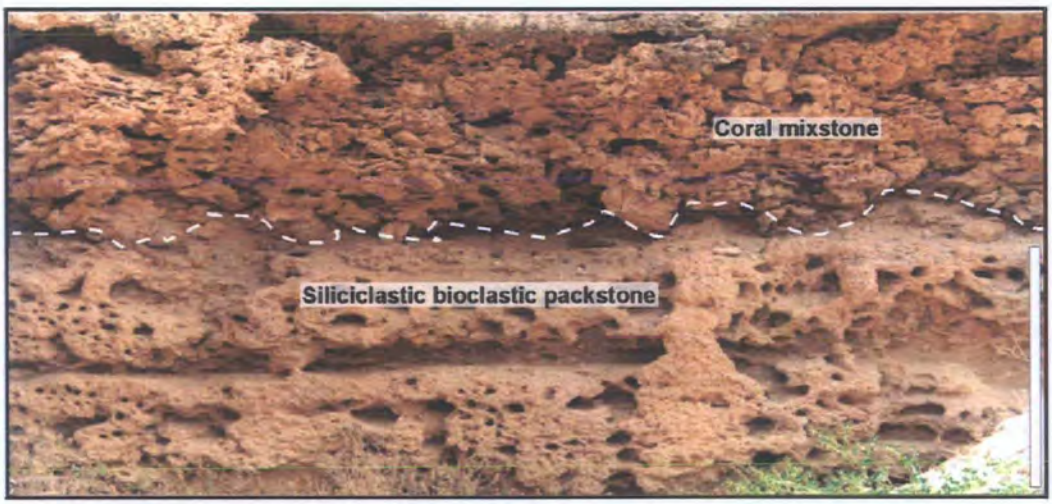


Figure 5.16a The nature of the lower contact between the coral mixstone and siliciclastic bioclastic packstone facies (dashed line). The contact is irregular but non-erosional. At this locality, the coral mixstone is dominated by flat, thin platy coral colonies (Casa de Peñalver section, GR 61401296, facing S). Scalebar=1 m.

Figure 5.16b and **c** The erosive nature of the upper contact of the coral mixstone with the calcareous lithoclastic conglomerate facies (white dashed line). Along the northern ridge section (**b**, log FA-2, beds 8 and 9) the apparent depth of erosion is estimated to be at least 1 m. Equivalent depths are estimated for the carbonate interval exposed on the southern ridge (**c**, GR 61251318). The absence of conglomerate clasts within the reef framework indicates that coral development did not coexist with siliciclastic input. Layered conglomerates onlap the irregular upper surface of the coral mixstone (**c**), thus inferring that reef erosion occurred prior to burial by siliciclastics. See text for discussion.

facies is characterised by an abundance of *in situ* coral colonies that demonstrate platy, foliaceous and domal morphologies (**Figures 5.16a, 5.17a** and **b**). Colonies are tightly packed forming a framework. In addition to corals, the coral mixstone facies contains a moderately diverse fossil assemblage. Intact echinoids up to 7 cm in diameter and gastropods up to 2 cm are present in interskeletal areas.

Bioclastic pack/wackestone matrix occurs locally in interskeletal areas, and may have been recently eroded out (**Figures 5.16a** and **5.17a**). The matrix is dominated by dark brown micrite (56.5 %) with minor silt-grade quartz (<0.5 %) and quartz-mica schist (<0.5 %). This facies contains a diverse biota that comprises bryozoa (2 %), echinoids (1 %), bivalves (5 %), coralline algae (2 %), victoriellids (0.5 %), small benthic foraminifera (1 %), planktonic foraminifera (1.5 %) and *Amphistegina* (<0.5 %). Most fossils are fragemented and show evidence of abrasion. Gastropods and echinoids are often preserved intact. Bivalves are dominated by pectens with subordinate oysters. Fragments are up to 2 cm across.

Locally, and particularly at the Casa de Peñalver section (**Figure 5.1**), the upper surfaces of the coral mixstone facies are characterised by columns of concentrically laminated carbonate (**Figures 5.17c** and **d**). Laminae are around 1 mm thick. Measured column width and height is up to 15 cm. Fine-grained carbonate material may be preserved between laminae. The upper surfaces of columns are planar and commonly demonstrate small, cyclindrical borings (**Figure 5.17d**).

Diagenesis

Intergranular drusy and equant calcite spar cement occurs around bioclasts and siliciclastic grains. All originally aragonitic bioclasts (including corals, bivalves and gastropods) have undergone dissolution and mouldic pores have been wholly, or partially, cemented with drusy calcite spar. Very high inter-skeletal macroporosity (up to 50 %) has developed with the preferential weathering of the soft wack/packstone matrix (**Figures 5.16a** and **5.17a**).

Interpretation: environment of deposition

The coral mixstone facies contains a marine faunal assemblage. Deposition within the photic zone under normal open marine conditions is inferred from the abundance of stenohaline biota. Agitated conditions are suggested from the presence

Figure 5.17a Massive and foliaceous coral colonies (log FA-12, bed 9). The matrix has preferentially weathered. Scalebar=1m.

Figure 5.17b Domal coral colony (*Goniastrea?*) within the coral mixstone facies (Casa de Peñalver section). Scalebar=2 cm.

Figure 5.17c Columnar stromatolites observed on the upper surface of the coral mixstone facies (Casa de Penalver, GR 61501292). Individual columns are up to 15 cm in height, although erosion is inferred from the planar surfaces. Original microbial structures have been replaced with calcite, hence the white crystalline appearance of columns. Scalebar=50 cm.

Figure 5.17d Detailed view of the columnar stromatolites pictured above. Columns are composed of numerable tightly concentric calcite laminae. The upper surfaces are often planar, inferred to be a consequence of erosion. Note the abundant small, cyclindrical borings on the upper surfaces (arrowed). Scalebar=50 cm.

of fragmented and abraded bioclasts in the matrix, although broadly low-energy conditions are suggested from the abundance of micrite. It is postulated that the accumulation of micrite would have been encouraged by the dense coral growth (Insalaco 1998).

The concentrically laminated carbonate columns present on the upper surfaces of this facies are interpreted as microbial in origin, although no organic microbial structures have been preserved. Using the classification scheme provided by Riding (2000), these structures are classified as columnar stromatolites. There is no evidence of microbial/stromatolite growth within the irregular coral framework. It is concluded therefore that stromatolite growth occurred post-reef.

Modern stromatolite/microbialite development tends to occur in intertidal environments (Tucker and Wright 1990). In the Bahamas, columnar stromatolites occur in subtidal, lagoonal environments (Reid *et al.* 1999). An abundance of microbial structures has been attributed to the limited grazing by invertebrates, perhaps as a consequence of enhanced salinities/environmental degradation (Kendall and Skipwith 1968, Garrett 1970 and Riding *et al.* 1991). Riding (2000) states that although biologically stimulated, the over-riding environmental control on stromatolite development is seawater chemistry and the supersaturation state of seawater with respect to carbonate minerals. Detailing the changing seawater chemistry was beyond the scope of this study and thus this hypothesis cannot be tested. A significant change in environmental conditions towards the top of the coral mixstone facies is inferred from the transition from a diverse stenohaline assemblage to a limited, microbial-dominated assemblage.

In summary, the coral mixstone facies is interpreted as an *in situ* coral build-up that developed in moderate depths in a normal marine environment with a low siliciclastic influx. Towards the closing stages of coral growth, environmental perturbations such as unfavourable changes in seawater chemistry, decrease in abundance of grazers etc, resulted in the development of columnar stromatolites. Comparable mixstone coral build-ups are described from a number of the Upper Miocene Mediterranean basins. The coral mixstone facies is interpreted as an equivalent of the 'Thicket Zone' of the Sorbas Basin, as described by Braga and Martin (1996). In addition, irregular laminar coral growth forms are described from the Reef Core facies of the Melilla-Nador Basin, Morrocco (Saint Martin and Cornée

1996), and the reef slopes of the Llucmajor Area, Mallorca (Pomar *et al.* 1996). Comparable stromatolitic facies are described from the Sorbas Basin (Riding *et al.* 1991), although the columnar stromatolites are more typically associated with coarse siliciclastics rather than corals.

5.2.3.3 Laminated coral pillarstone

Lithologies: Coral pillarstone

Occurrence and bed characteristics

This facies is only encountered at the margins of the carbonate units in the northern ridge. The laminated coral pillarstone facies can only be traced laterally for 5 metres. This facies is not encountered on the southern ridge or in the Casa de Peñalver section. The lower contact of this facies with the laminated coral platestone is gradational. The upper contact with the calcareous lithoclastic conglomerate facies is sharp and locally erosional.

Lithological description

Exposed surfaces of this facies weather to a pale grey to light brown colour. Fresh surfaces are white and have a finely crystalline texture. The laminated coral pillarstone facies is characterised by an abundance of dissolved stick-like branching coral colonies with concentrically laminated calcite crusts up to 1.5 cm thick (**Figures 5.18a** and **b**). Calcitic laminae are less than 1 mm thick, quite poorly defined and have a crystalline texture (**Figure 5.18b**). Coarse calcareous litharenitic sand with well-rounded lithic pebbles is preserved between coral colonies in the upper 50 cm of this facies (**Figure 5.18b**). Siliciclastic grains are up to 1.5 cm in diameter. Grains are dominated by metasediments (up to 30 %), limestone (up to 5 %) and intraclasts (up to 2 %) floating within a coarse-sand matrix. Grains are well-rounded, moderate to poorly sorted and sub-spherical (**Figure 5.18b**).

The matrix within the lower 50 cm of this facies is dominated by micrite (12 to 30 %) with minor silt-grade quartz (0.5 %) and quartz-mica schist (0.5 %). Biota include gastropods (1 to 2 %), bivalves (5 %), echinoids (0.5 to 5%), bryozoa (0.5 to 2 %), victoriellids (< 0.5 %), miliolids (0.5 %), *Amphistegina* (< 0.5 %), victoriellids (< 0.5 %), planktonic foraminifera (< 0.5 %) and small hyaline benthic foraminifera

Figure 5.18a Field view of the laminated coral pillarstone facies. Stick-like corals (arrowed) have been completely dissolved out. This facies occurs only at the margins of carbonate bodies. Scalebar=5 cm.

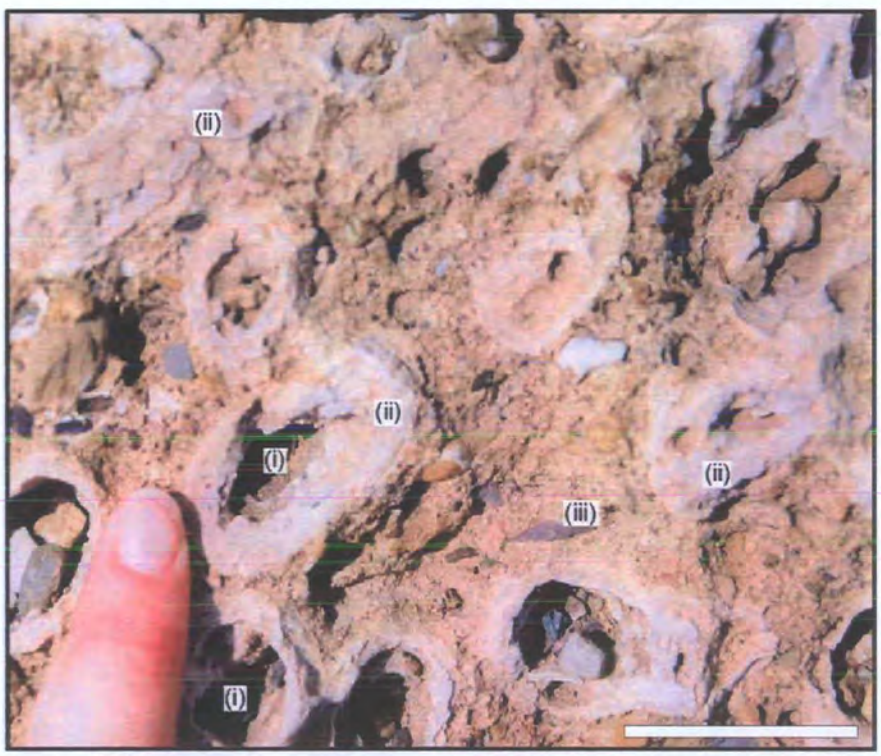


Figure 5.18b Close-up of above. Vertical stick-like corals have been completely leached (i), although the outline of branches has been preserved by concentric, laminar microbial structures (ii). Coarse sand-grade siliciclastic material (iii) occurs between coral branches. Scalebar=2cm.

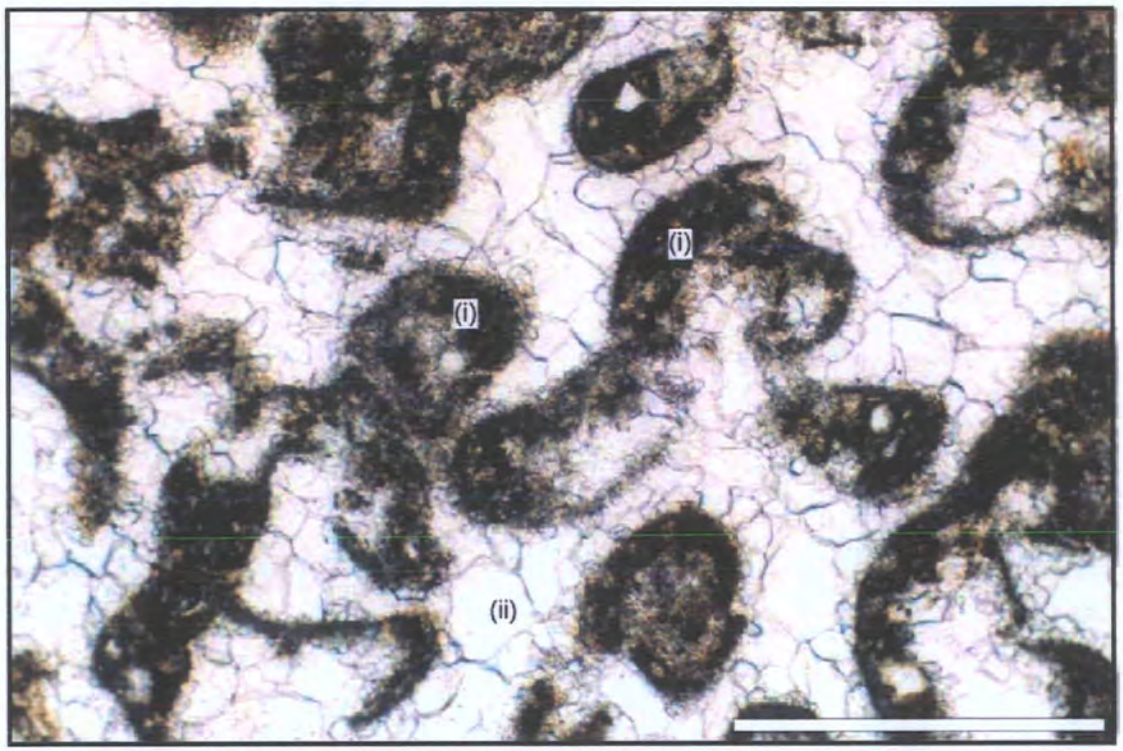


Figure 5.19a Photomicrograph of the laminated coral pillarstone facies. Clotted micritic textures (i) are interpreted to be microbial in origin. The original coral skeleton has been leached, with the resultant mouldic pores completely occluded with drusy calcite cement (ii) (sample LFA 75). Scalebar=5 cm.

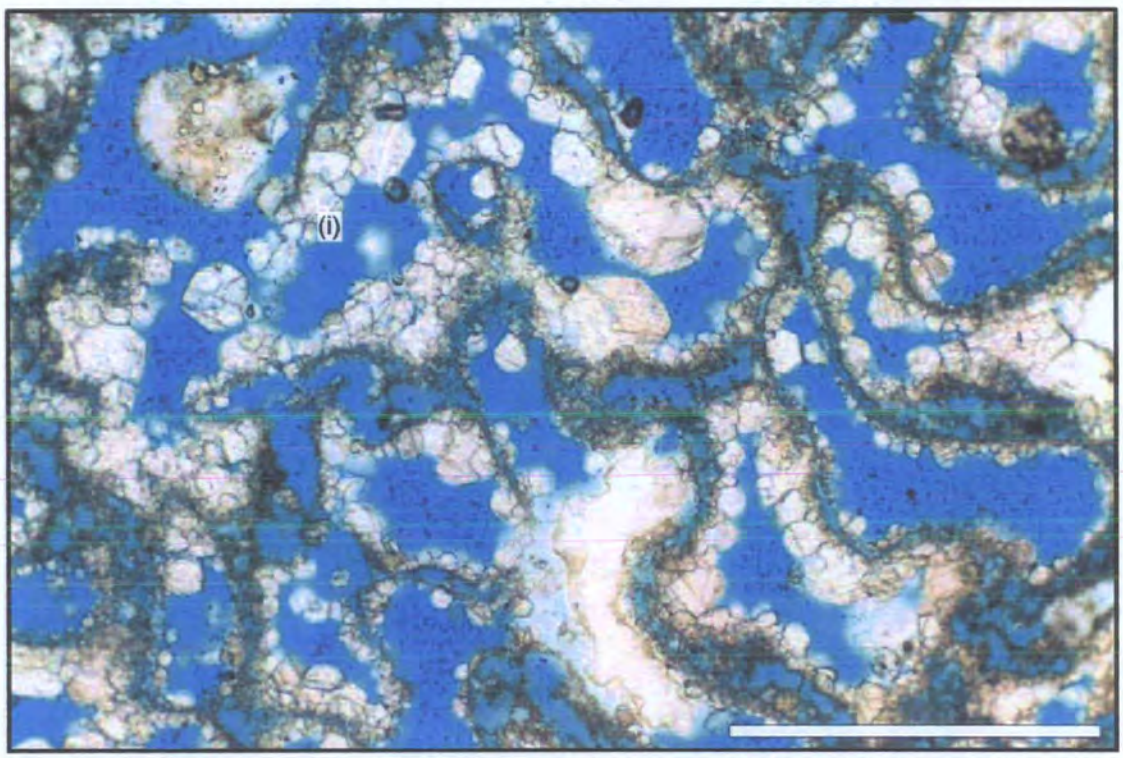


Figure 5.19b Photomicrograph of the laminated coral pillarstone facies. Originally aragonitic coral skeletons have been completely dissolved. Mouldic pores have been partially occluded with blocky calcite rim cements (i), although significant porosity remains in this sample (sample LFA 19b, log FA-3 bed 5) Scalebar=2cm.

(< 0.5 %). Most fossils are fragmented with evidence of abrasion. Fragments are up to 2 cm across.

Diagenesis

Originally aragonitic bioclasts (corals and bivalves and gastropods) have been dissolved out and biomouldic pores wholly, or partially, infilled with drusy and equant calcite spar cement (**Figures 5.19a** and **b**). Clotted, micritic textures may be associated with corals in this facies (**Figure 5.19a**). Twenty to 40 % of the laminated coral pillarstone facies is vuggy, biomouldic and intergranular porosity (**Figure 5.19b**).

Interpretation: environment of deposition

The laminated coral pillarstone facies contains a marine faunal assemblage. Deposition within the photic zone under normal marine conditions is concluded from the abundance of stenohaline biota. Branching coral morphologies characteristic of this facies are consistent with Miocene examples of coral occurrences in shallow water depths in areas of active (siliciclastic) sedimentation (Santisteban and Taberner 1988, Pomar *et al.* 1996). The coral *Porites,* dominant in the Mediterranean during the late Miocene, is well known for its ability to withstand high levels of (fine-grained) siliciclastic input (Hubbard and Pocock 1972). It has been suggested that the stick morphology responds to a need for fast growth rates (to compete for space, to avoid burial and to maximise light availability) (Grasso and Pedley 1988, 1989, Saint Martin 1990). Intuitively, the relatively narrow *Porites* branches would be fragile and high-energy environments, or environments experiencing influxes of coarse siliciclastic material, would quickly result in fragmentation of colonies. It is suggested by Esteban *et al.* (1977) and Esteban (1979) however, that sediment packed around branches effectively supported them. Agitated conditions are inferred from the presence of abraded and fragmented bioclasts in the matrix and proximity to a siliciclastic source is inferred by the occurrence of silt to pebble-grade lithic and detrital quartz clasts. Some transport and reworking of siliciclastic grains prior to deposition and lithification is suggested from the rounded nature of lithic grains in the matrix.

The laminated calcitic crusts present around often dissolved-out coral branches are interpreted as microbial in origin although no primary cellular microbial structures are preserved. Using the classification scheme of Riding (2000), these structures are classified as stromatolites. In thin section, clotted micritic fabrics associated with coral skeletons are also interpreted to be microbial in origin. Using the classification scheme of Riding (2000) these textures would be classified as thrombolites. Modern microbial structures are commonly associated with tidal flat and inter-tidal environments (Tucker and Wright 1990) although a sub-tidal environment of deposition is favoured due to their association with *in situ* stenohaline biota, and the absence of features that could indicate subaerial exposure.

Kendall and Skipwith (1968), and Riding *et al.* (1991) have attributed an abundance of microbial structures to limited grazing by invertebrates, perhaps as a consequence of enhanced salinities/environmental degradation. Normal open marine conditions are interpreted for this facies. Nutrient availability (and excess) is a potential control on the development of microbialites (Reitner and Neuweiler 1995). A coral reef framework with cavitites and vertically dipping substrates (i.e. vertical branches) may favour microbialite accretion by providing protection from grazers such as gastropods (Riding *et al.* 1999).

Determining the relative timing of microbial encrustation on corals is difficult. It has been suggested by Martín *et al.* (2000) that microbial crust development can accompany coral growth in normal marine conditions with apparently no detrimental effects. It has been suggested that stromatolite accretion could occur at the base of actively growing branching coral colonies (Riding, pers. comm. 2002). Microbial coatings may have a beneficial role in strengthening coral reef frameworks in high-energy marginal marine environments, particular those that are characteristic of the Mediterranean in the late Miocene (Riding *et al.* 1991 and Martín *et al.* 2000).

In summary, this facies is interpreted as branching coral development with contemporaneous microbial encrustation in a moderate to high-energy, shallow marine environment with a coarse, siliciclastic influx in the later stages of its development. The laminated coral pillarstone facies is interpreted to be the equivalent of the branching *Porites* and *Tarbellastraea* zone of the coral-microbial bioherms of the Sorbas Basin, as described by Riding *et al.* (1991) and Martín *et al.* (1998). In addition, microbial crusts on branching corals are documented in Messinian reefs of

the San Miguel de Salinas Basin (Reinhold 1995). Microbial crusts of the San Miguel de Salinas Basin are however biotically more diverse than those observed in the Altorreal section (Reinhold 1999). In addition crusts, which may be up to 1 cm thick, comprise coralline algae (*Lithothamnion sp.*) and encrusting (acervulinid) foraminifera (Reinhold 1999).

5.3 Altorreal Facies Associations and Summary

Excluding the silty marl, the facies described in **Section 5.2** were all deposited in a moderate to high-energy, shallow-marine (subtidal) environment proximal to the coastline. Characteristics of modern-day equivalent systems, and Miocene equivalents from SE Spain and the Mediterranean have been used to aid interpretation of these facies associations. A depositional model of the Altorreal area may be formulated by understanding the three-dimensional facies architecture. The three-dimensional facies architecture has been reconstructed from the lateral correlation of graphic logs (**Figures 5.20** and **5.21**). Numbered sections are roughly parallel and perpendicular to the dip (and progradation) direction of units.

5.3.1 Facies associations

The identified facies of the Altorreal section occur in three associations summarised in **Table 5.2.** Facies are not exclusive to an association. A summary log through the Altorreal section illustrating the diagnostic characteristics of identified facies associations and the stratigraphic arrangement of associations is presented on **Figure 5.22.**

5.3.1.1 Facies association 1

Facies: Calcareous lithoclastic conglomerate
Calcareous litharenite
Siliciclastic bioclastic conglomerate
Silty marl

Characteristics

Siliciclastic sediments comprise the majority of studied sections within the Altorreal study area (**Figures 5.2** and **5.3**). The calcareous lithoclastic conglomerate, with intercalations of the calcareous litharenite and siliciclastic bioclastic packstone

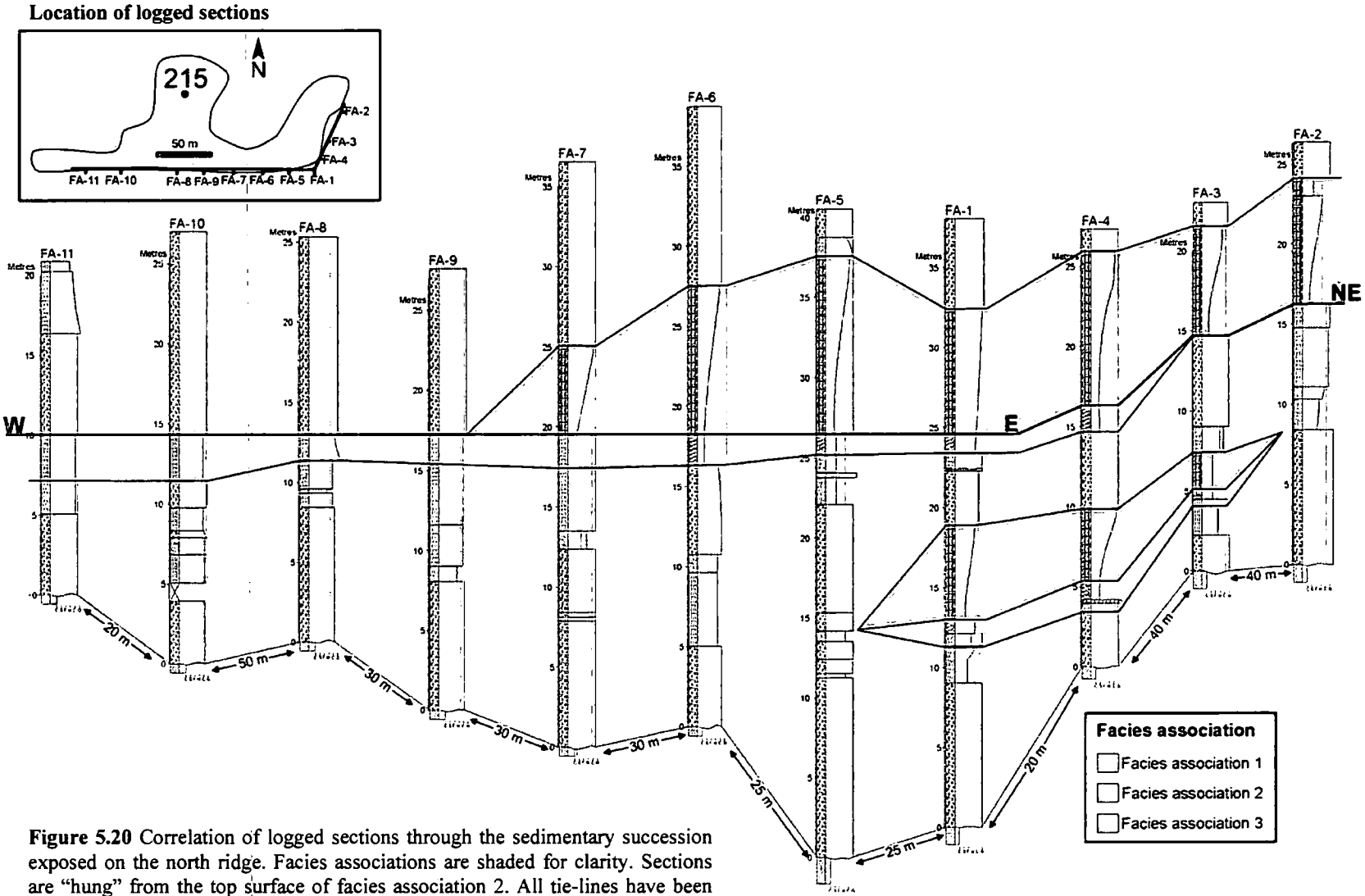


Figure 5.20 Correlation of logged sections through the sedimentary succession exposed on the north ridge. Facies associations are shaded for clarity. Sections are "hung" from the top surface of facies association 2. All tie-lines have been traced in the field.

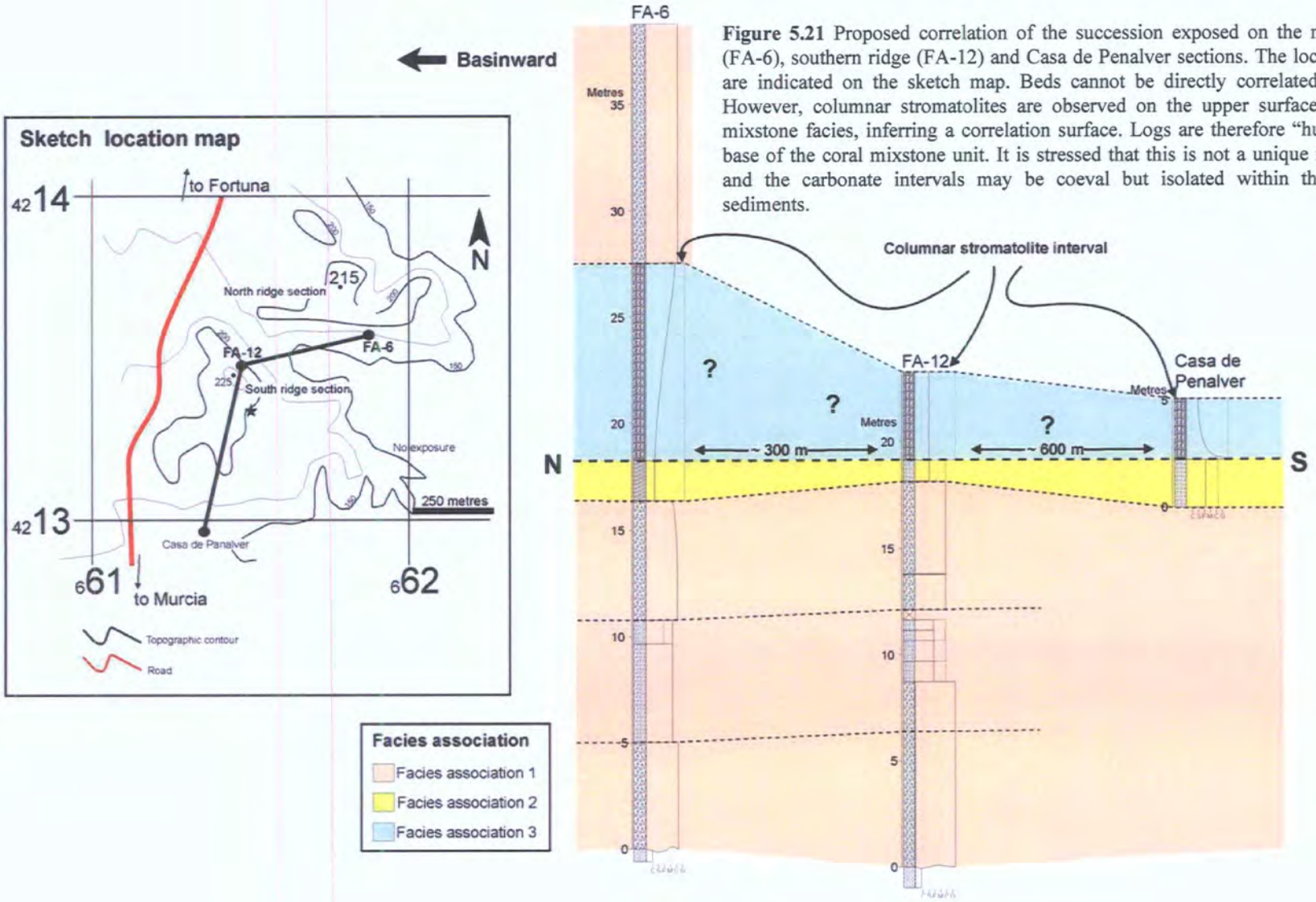


Figure 5.21 Proposed correlation of the succession exposed on the northern ridge (FA-6), southern ridge (FA-12) and Casa de Penalver sections. The locations of logs are indicated on the sketch map. Beds cannot be directly correlated in the field. However, columnar stromatolites are observed on the upper surface of the coral mixstone facies, inferring a correlation surface. Logs are therefore "hung" from the base of the coral mixstone unit. It is stressed that this is not a unique interpretation, and the carbonate intervals may be coeval but isolated within the siliciclastic sediments.

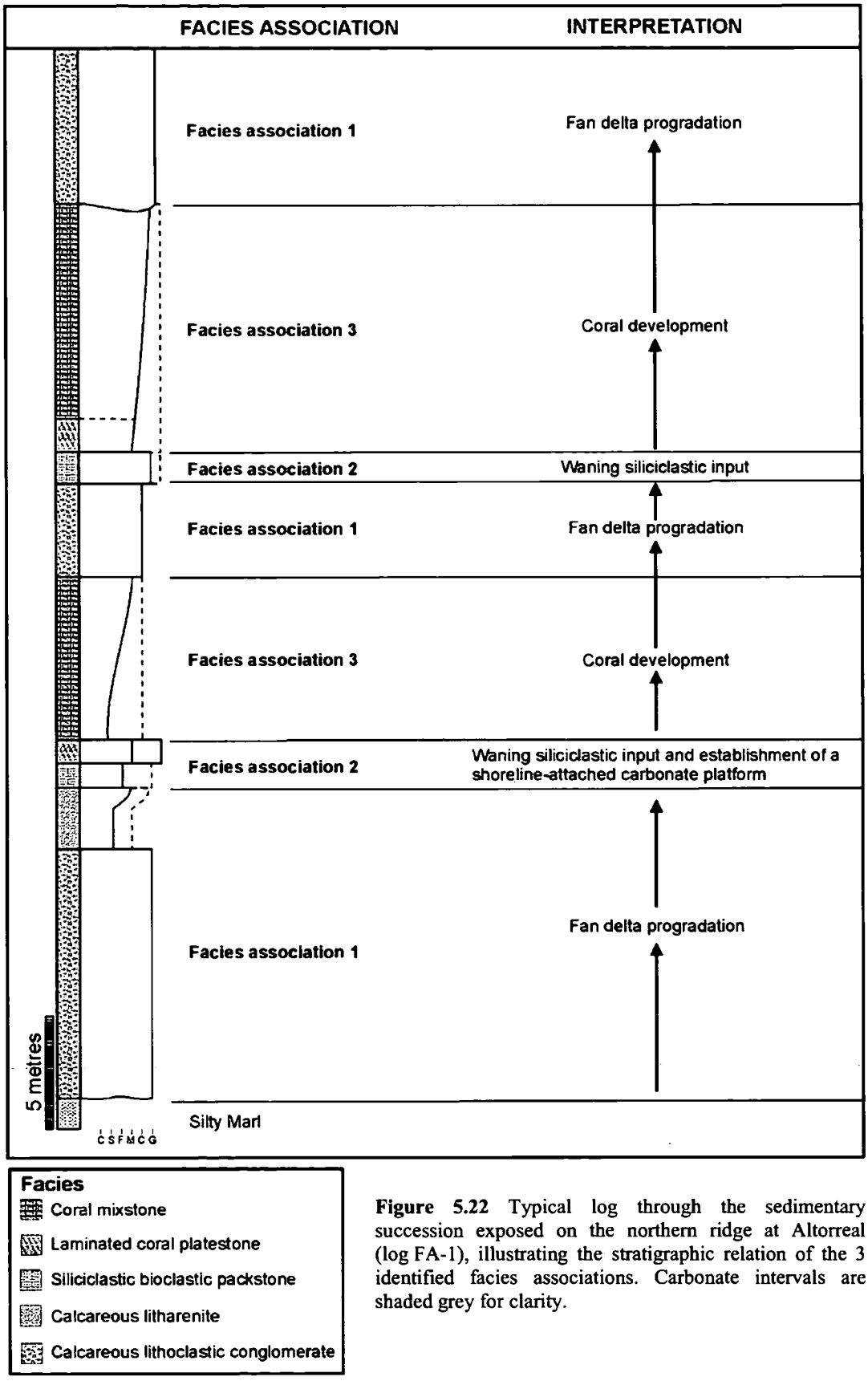


Figure 5.22 Typical log through the sedimentary succession exposed on the northern ridge at Altorreal (log FA-1), illustrating the stratigraphic relation of the 3 identified facies associations. Carbonate intervals are shaded grey for clarity.

facies, characterise this association. The association is typically 10 to 15 m thick (**Figure 5.2**). The fauna present is marine, and consists of robust oysters, pectens, gastropods, echinoids, serpulids, *Amphistegina,* small rotaline and planktonic foraminifera. Most fossils are fragmented, and fragments are angular and poorly sorted. Relatively fine-grained units within this association may be bioturbated and large boulders may be encrusted by oysters or bored by lithophagids.

Bioclastic conglomerates and sands of this association are observed to directly overlie the silty marl facies (**Figures 5.4a** and **5.22**). The contact is erosional, sharp and undulating. Lenses of silty marl within the calcareous lithoclastic conglomerate facies are interpreted as large rip-up clasts (**Figure 5.4a**).

Beds within this association are arranged as mega-forsets (clinoforms), and show vertical increases in inclination over distances of less than 10 metres. Bottom-sets, fore-sets and occasionally top-sets are preserved (**Figure 5.2**). Fore-sets have progradational geometries (**Figure 5.2**). Measured foreset orientation shows two different progradation directions (**Figure 5.24**). The lower conglomerates that directly overlie the silty marl facies show a northwest progradation direction. Conglomerates from the top of the exposed succession that overlie the upper carbonate interval demonstrate a west to southwest progradation direction (**Figure 5.24**). Conglomerates exposed within a valley to the north of the study area demonstrate a southeast progradation direction (**Figure 5.24**).

Within the clinoforms, the calcareous lithoclastic conglomerate facies is observed to pass down-dip into interbedded pebbly calcareous litharenites and siliciclastic packstones (**Figures 5.2** and **5.20**). Deposition is interpreted to have been intermittent and from debris-flows (**Section 5.2.1.1**). A schematic representation of this facies association is presented in **Figure 5.23a**.

Interpretation

Facies association 1 is interpreted to represent the deposition of coarse-grained material in high-energy marine fan delta environment proximal to the siliciclastic sediment source. A dominant progradational sedimentary regime is inferred from the cross-stratified nature of beds within this association. It is inferred from the apparent juxtaposition of high and low-energy facies at the base of this association

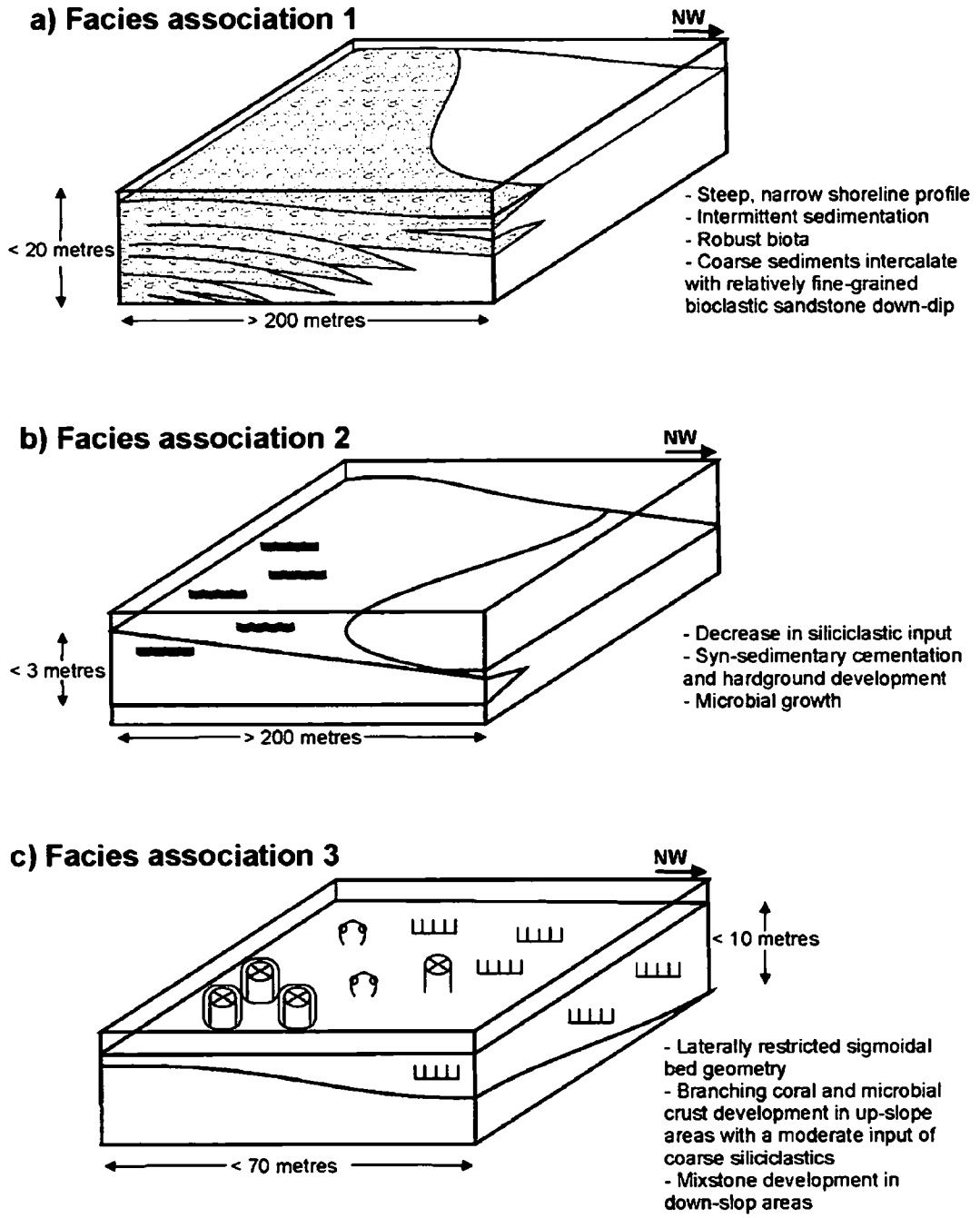


Figure 5.23 Altorreal facies associations. See text for discussion.

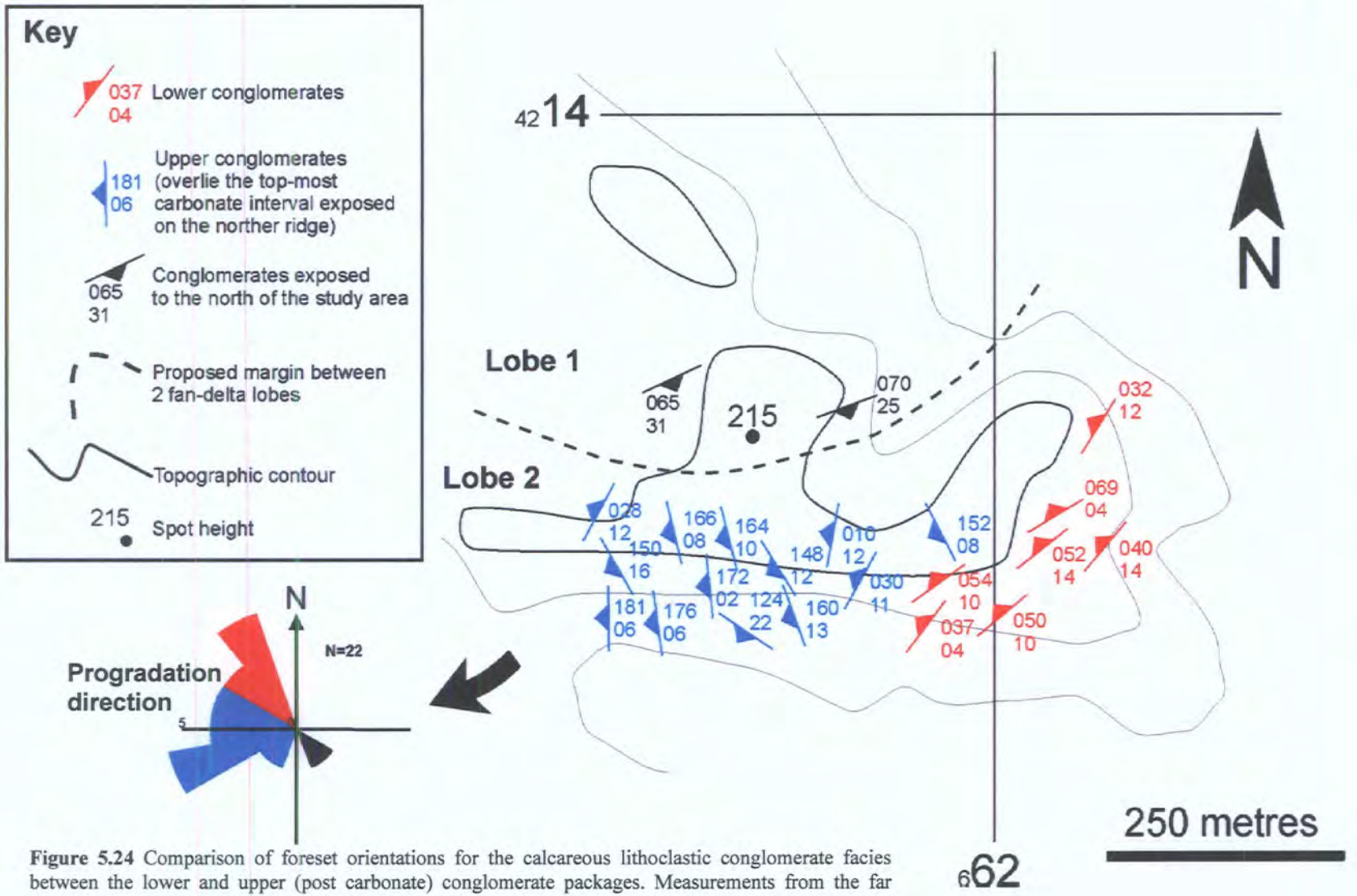


Figure 5.24 Comparison of foreset orientations for the calcareous lithoclastic conglomerate facies between the lower and upper (post carbonate) conglomerate packages. Measurements from the far north of the study area (black) are also illustrated. The postulated margin of two different fan delta lobes is outlined (dashed line).

(**Figure 5.4a**), that the high-energy delta system with locally very steep gradients was prograding directly onto sediments of a deeper, prodelta-type setting. This situation is characteristic of Gilbert-type delta systems (Postma 1990). Facies association 1 represents the proximal fore-sets of a fan-delta, dominated by debris flows. Water depths are estimated to have been around 10 to 15 m, based upon the preservation of bottom-, fore- and occasionally topsets. In comparison to other Gilbert-type delta environments described from the Miocene basins of SE Spain (i.e. the Granada Basin, Braga *et al.* 1990), there is no evidence for prolonged sub-aerial exposure (i.e. mature soil development) and erosion although the potential for minor early pedogenesis is discussed in **Section 5.2.1.1**.

The progradation direction of upper and lower carbonate packages differs slightly (**Figure 5.24**), although the composition of clasts does not change, inferring that there was no change in sediment provenance. A broad west-southwest to northwest progradation direction is concluded for the fan delta system. The progradation direction of conglomerates to the north of the study is significantly different (**Figure 5.24**), thus it is suggested that these sediments are deposits of an unconnected lobe.

Facies association 1 is comparable to the siliciclastic dominated (sd) facies of Wrobel and Michelzik (1999), also interpreted as proximal fan-delta deposits. In addition, Braga *et al.* (1990) describe a number of fan-delta megasequences, consisting of silty marls followed by silts and sands and then sands and conglomerates and finally conglomerates. These sequences, representing evolution from a deep marine to shallow marine environment, are typically topped off by red palaeosols, which represent continental emergence.

5.3.1.2 Facies association 2

Facies: Siliciclastic bioclastic packstone
Laminated coral platestone

Characteristics

This association vertically succeeds facies association 1 (**Figure 5.22**). The mixed carbonate-siliciclastic association is 2 to 3 metres in thickness and contains beds of the siliciclastic bioclastic packstone and laminated platestone facies. The

siliciclastic bioclastic packstone facies grades laterally into the laminated platestone facies (**Figure 5.2**).

This facies association is characterised by an open marine fauna including oysters, pectens, coralline algae, gastropods, bryozoa, corals and small rotaline benthic foraminifera. Beds of this facies association are characterised by a relatively low siliciclastic influx with sediment reworking, probably by marine currents and wave action. These beds are characterised by early syn-sedimentary cementation (i.e. hardground development) accompanied by boring and encrustation by marine organisms, and stromatolite mat development (**Sections 5.2.2.2** and **5.2.3.1**).

Interpretation

Facies association 2 is interpreted to represent a waning siliciclastic influx and the establishment of a low-angle, shoreline-attached carbonate platform environment. Negligible siliciclastic influx and low accumulation rates resulted in hardground development. It is suggested that stromatolite development may have been influenced/favoured by syn-sedimentary cementation (Riding pers. comm. 2002), which would be typical of shallow marine environments with high current activity and/or reduced sedimentation.

The evolution from facies associations 1 to 2 is interpreted to represent a change from a progradational to retrogradational and aggradational sedimentary regime (**Figures 5.22** and **5.23b**). A similar facies succession is described and interpreted in the same way from the Lorca Basin (Wrobel and Michelzik 1999).

5.3.1.3 Facies association 3

Facies: Coral mixstone
Laminated coral pillarstone

Characteristics

Facies association 3 may succeed facies association 2 or facies association 1 (**Figures 5.20** and **5.22**). The carbonate facies association contains the coral mixstone and laminated coral pillarstone facies (**Figure 5.22**). This association forms laterally restricted sigmoidal carbonate units that extend laterally (perpendicular to the siliciclastic progradation direction) up to 70 m (**Figures 5.2** and **5.14**). The lateral extent of this facies parallel to the dip direction is estimated to be up to 750 m,

although quarrying of the section has made correlation of beds difficult as stratigraphic contacts are not continuously exposed (**Figure 5.21**). Vertical thickness is up to 10 m. The carbonate association thickens down-dip from 0 to 3 m where the laminated platestone and laminated coral pillarstone dominates, to over 10 m where the coral mixstone dominates (**Figure 5.14**).

The carbonate association is dominated by *in situ* corals and fragmented reef-associated fauna including echinoids, gastropods, pectens, coralline algae and bryozoa. The corals in the carbonate facies association have a very low diversity and consist almost entirely of *Porites*. Time-equivalent reefs in the Fortuna Basin also contain *Tarbellastrea*, *Monastrea* and *Siderastrea* (Santisteban 1981, Santisteban and Taberner 1988).

Interpretation

The carbonate facies association is interpreted as an *in situ* reef build-up representing further substrate colonisation by corals, in addition to pre-existing microbial mats, during a period of negligible siliciclastic input. It is postulated that the sigmoidal morphology of the carbonate units (and facies association 3) reflects the morphology of the siliciclastic substrate.

Relative depositional depths for each part of the carbonate unit have been determined using coral colony morphologies, as schematically illustrated on **Figure 5.23c**. The shallowest water conditions occurred where the carbonate units pinch-out and are dominated by vertical stick-like branching coral colonies. The coral mixstone facies developed down-dip in relatively deeper water conditions unaffected by siliciclastic input. It is postulated that this facies formed thicker units as a response to higher accommodation space (and deeper water conditions). It is interpreted that the facies association 3 represents a predominantly aggradational sedimentary regime, with a relatively minor progradational element (as shown by the bedding in the southern ridge section, **Figure 5.3**).

The nature of microbialite development varied. Development in very shallow water with a moderate siliciclastic input is represented by the laminated coral pillarstone facies. The absence of microbialite development in the coral mixstone facies indicates microbial development only occurred in very shallow water environments. Columnar stromatolite development was not contemporaneous with

coral growth (**Section 5.2.3.2**). Columnar stromatolite development is related to environmental degradation and the loss of stenohaline biota (including corals). Columnar stromatolites are common within low-energy, shallow marine lagoonal settings, often with slightly enhanced salinities (Reid *et al.* 1999, Pomar 2001). A significant change in the conditions of the depositional environment is inferred from the development of columnar stromatolites.

5.3.1.3.1 The carbonate-siliciclastic boundary

In all studied sections of the Altorreal study area, facies association 1 always succeeds facies association 3. The contact is sharp and erosional (**Figure 5.16b** and **c**). Clinoforms of facies association 1 are observed to onlap the carbonate bodies (**Figure 5.16b**). The depth of erosion is typically around 1 m. Carbonate fragments are relatively rare in the overlying conglomerates although boulder-grade fragments, inferred to be locally derived, are observed (**Figure 5.5c**).

In conclusion, the boundary is interpreted as a sharp change from a carbonate-dominated, aggradational sedimentary regime characterised by *in situ* coral growth to a siliciclastic dominated, progradational sedimentary regime characterised by conglomeratic and coarse litharenite deposition. A period of exposure and erosion of the reef carbonate units is interpreted prior to burial by siliciclastics. This is supported from the development of columnar stromatolites that often represent a significant period of non-deposition (Hubbard, pers. comm. 2003). It is suggested that reefs had undergone cementation prior to burial from the rarity of carbonate fragments in the overlying siliciclastics. Rapid burial by siliciclastics would also have precluded reef destruction.

5.3.2 Depositional model

Using the facies characteristics and interpretations described in **Sections 5.2**, in combination with the lateral sedimentary log correlations illustrated on **Figures 5.14** and **5.21**), a three-dimensional evolutionary model of the main ridge section of the Altorreal depositional succession is proposed (**Figure 5.25**).

Stage 1: Marl deposition

This initial stage is characterised by widespread deposition of the silty marl facies in a low-energy, open marine environment below storm-wave base (**Figure 5.25a**). Depth of deposition was in the order of tens of metres. Based upon the available stratigraphic literature for the Fortuna Basin (reviewed in **Chapter 4**), marl deposition was likely to have been a basin-wide event, although this cannot be determined using data from the Altorreal study area alone.

Stage 2: Fan-delta progradation

The second depositional stage represents the advance of a Gilbert-type fan-delta system into the low-energy silty marl environment. This high-energy incursion is characterised by erosional lower contacts with the underlying silty marl facies, and the presence of large silty marl rip-up clasts in the lower portions of the fan-delta (**Figure 5.4a**). There is no evidence for lobe margins or change in grain composition, and it is therefore interpreted that this lower fan-delta package was deposited from a single lobe/sediment source. Sedimentation was episodic, and deposition was dominated by debris flow events. Beds have a clear progradational geometry, and northwest fan-delta progradation is concluded from palaeocurrent data (**Figure 5.24**).

Stage 3: Siliciclastic hiatus

This stage represents a significant reduction in terrigeneous siliciclastic influx and a change from siliciclastic to mixed carbonate-siliclastic sedimentation with the development of a carbonate ramp-type system, with low rates of sedimentation. The biota is dominated by oysters, pectens and gastropods, with subordinate coralline algae, bryozoa and corals. Stromatolite development occurred in relatively shoreward/shallow parts of the environment. The presence of robust, platy coral colonies indicates subtidal, normal marine conditions throughout deposition.

Stage 4: Coral reef colonisation and growth

The prolonged period of low sedimentation rates and substrate stabilisation allowed coral colonisation. Reef growth is apparently laterally restricted, evidenced by the preservation of small carbonate bodies, and absence of reef talus.

Stage 5: Fan-delta progradation resumes

Fan-delta progradation resumes, burying the earlier reef system. Fan-delta progradation is towards the northwest (**Figure 5.24**). Siliciclastic deposition was episodic and high-energy. Robust oysters and branching *Porites* colonies opportunistically developed on pebbles and foreset surfaces respectively during quiescent periods. Within down-slope areas where the inclination of beds is reduced, the relative proportion of sandstone and siliciclastic packstone increased (**Section 5.3.1.1**).

Stage 6: Siliciclastic hiatus

A significant reduction in siliciclastic influx and subsequent carbonate development occurred during this phase. High current activity combined with the high porosity of carbonate sediments and low sedimentation rates resulted in hardground development but only at the top of the unit. Patchy, contemporaneous stromatolite development occurred also occurred. The hardground, which was significantly bored by lithophagid bivalves, acted as a stable substrate to colonisers such as warty branching coralline algae, oysters and byrozoa and eventually provided a stable substrate allowing coral colonisation (Stage 7).

Stage 7: Carbonate factory re-established

The pre-existing fan-delta sediments dipped broadly towards the northwest (**Figure 5.24**). The upper carbonate package in the northern ridge section developed as a laterally restricted clinoform, with bed thickness increasing down-dip (**Figures 5.14** and **5.20**). The carbonate unit pinched out in relatively high-energy shallow water where vertical, stick-like coral colonies with stromatolitic coatings dominated (**Section 5.3.1.3**). Down-slope, corals filled the available accommodation space (**Figure 5.23c**). It is interpreted therefore that the inheritied morphology of the siliciclastics strongly influenced the morphology of carbonate units. It is concluced that carbonate development occurred either towards the edge of a lobe or in a submarine depression on the fan i.e. an abandoned channel. None of the reef packages identified can be seen to interdigitate with prodelta sediments. A relative absence of reef talus material typically associated with reef deposits is explained by the

significant syn-sedimentary cementation and microbial encrustation of corals, which ultimately strengthened the reef build-up.

A significant change in the depositional environment is inferred from the development of columnar stromatolites and the loss of stenohaline biota. Stromatolites are often related to periods of enhanced salinity, non-deposition and high nutrient input (accompanied by subtle changes in seawater chemistry) (**Section 5.2.3.2**).

Stage 8: Fan-delta progradation resumes

The re-establishment of a fan-delta system and resumed input of coarse terrigeneous siliciclastic material into the basin is inferred by the transition from the carbonate to the siliciclastic facies association. It is possible that the reef system may have undergone a brief period of exposure and erosion (**Figure 5.16**). Prolonged subaerial exposure was unlikely due to the absence of features associated with this process, such as karstification. Reefal carbonate were progressively on-lapped by the prograding fan-delta.

In summary, the sedimentary succession at Altorreal comprises repetitive phases of siliciclastic progradation, a sedimentary hiatus and carbonate development (**Figure 5.22**). During the progradational phase siliciclastic material was transferred into a low-energy environment via a gravel-rich Gilbert-type delta system. The sedimentary hiatus represents abandonment of the fan-delta and switch to a mixed carbonate-siliciclastic sedimentary regime. Finally, the reefal carbonates represent the aggradational phase.

The overall arrangement of facies in the Altorreal section is progradational, inferred from the clinoform bed morphologies and the slight basinward movement of separate carbonate packages (**Figure 5.20**). The locus of carbonate development varied little between phases exposed in the north ridge section. It is inferred that the morphology of carbonate units is inherited from the siliciclastic substrate.

5.3.3 Altorreal-evidence for salinity variations in the Fortuna Basin?

Traditionally, the occurrences of stromatolites/microbialites have been used to identify 'difficult' environmental conditions where processes such as colonisation, growth, competition and predation are low (Steneck *et al.* 1998). In the studied area, it is estimated that up to 50 % of reef bioconstructors in shallow waters were microbial. This appears to contradict traditional views of microbialite development because normal marine conditions are inferred throughout, due to the co-occurence of *in situ* stenohaline biota. It is apparent however that coral diversity in the Altorreal reefs is very low, and dominated by *Porites*. *Porites* is notable for being resistant to wide salinity variations (19-45 %) (Braithwaite 1971, Downing 1985). However, most coral reefs in the Mediterranean underwent radical reductions in diversity towards the Late Miocene, and particularly around the Late Tortonian-Messinan (Esteban 1996) thus in this respect, the low diversity of the Altorreal reefs may be a response to a more regional factor.

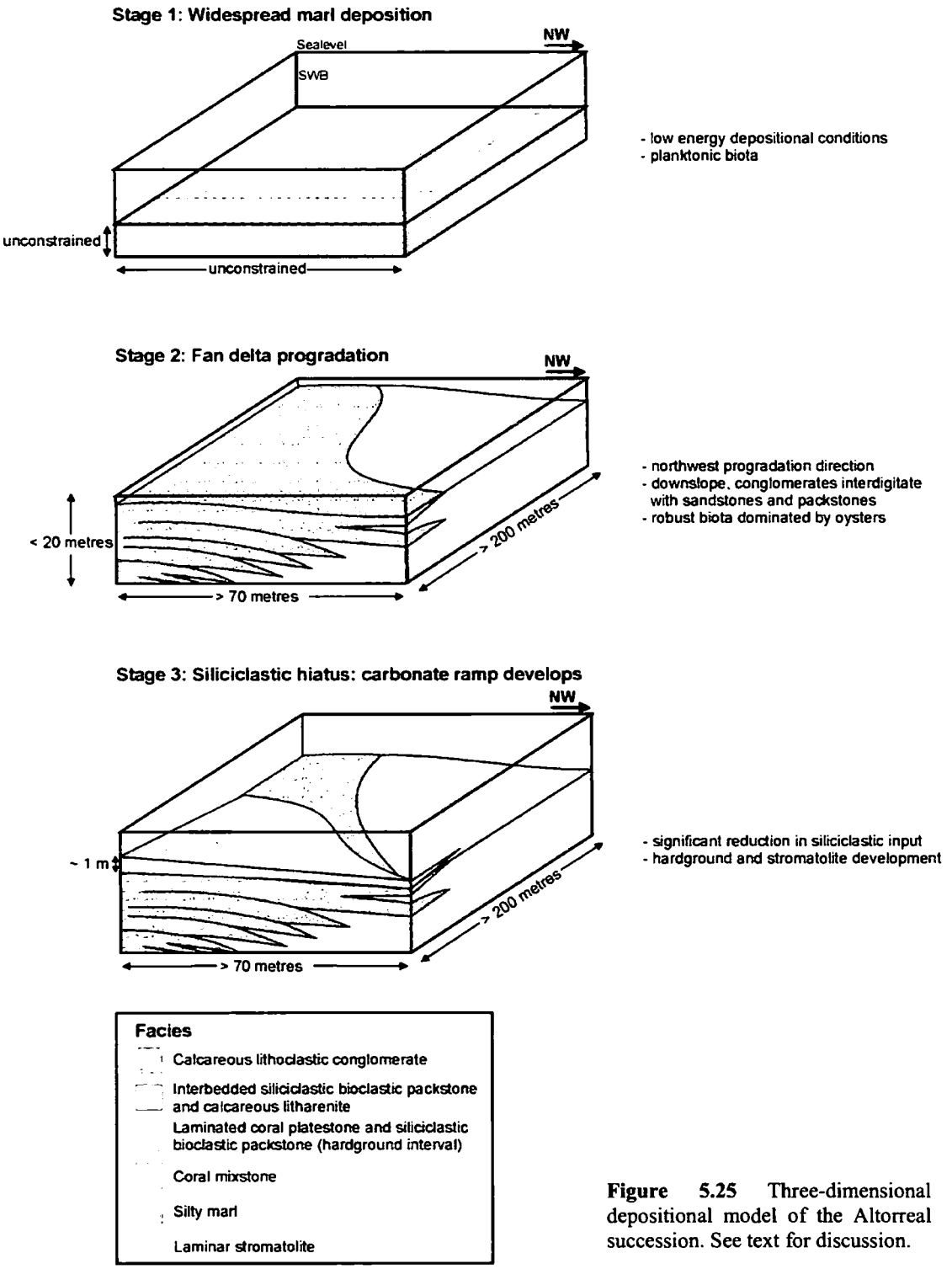
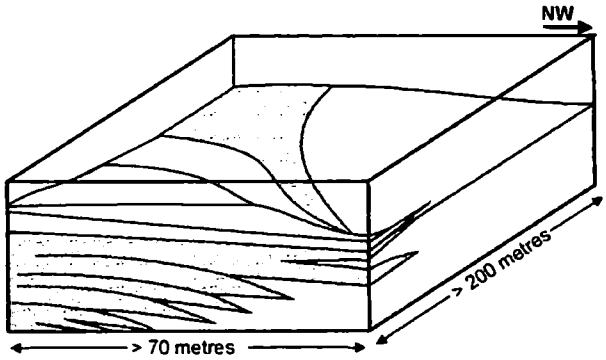


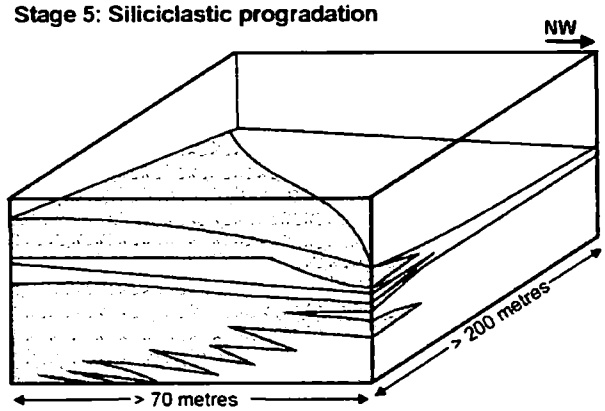
Figure 5.25 Three-dimensional depositional model of the Altorreal succession. See text for discussion.

Stage 4: Siliciclastic hiatus: coral development



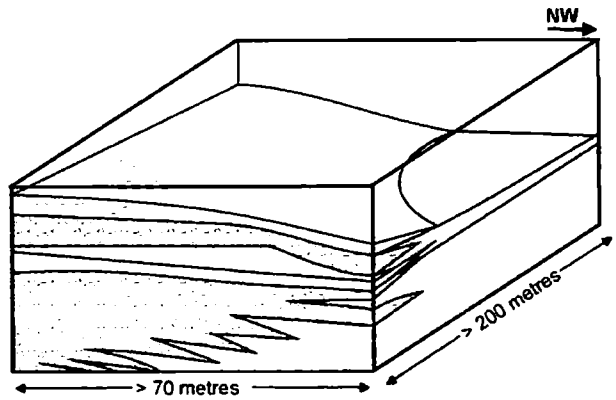
- carbonate develops as a clinoform, thickening downslope
- branching corals develop in the shallowest areas, with platy, foliaceous and domal forms present downslope

Stage 5: Siliciclastic progradation



- resumed northwest progradation of the fan delta, burying the reef

Stage 6: Siliciclastic hiatus: carbonate ramp develops



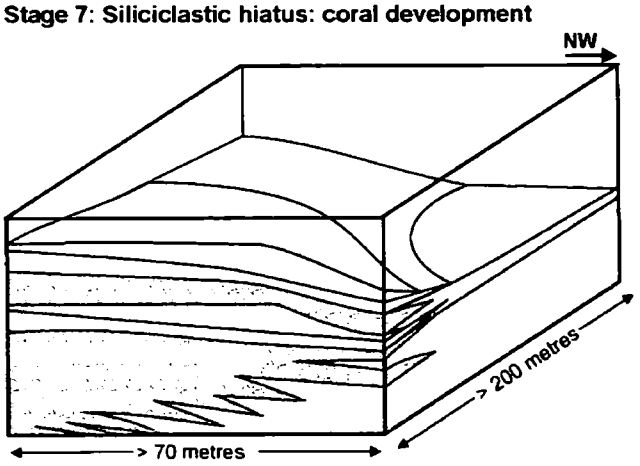
- hardground development with laminar stromatolites in relatively upslope areas

Facies

- Calcareous lithoclastic conglomerate
- Interbedded siliciclastic bioclastic packstone and calcareous litharenite
- Laminated coral platestone and siliciclastic bioclastic packstone (hardground interval)
- Coral mixstone
- Silty marl
- Laminar stromatolite

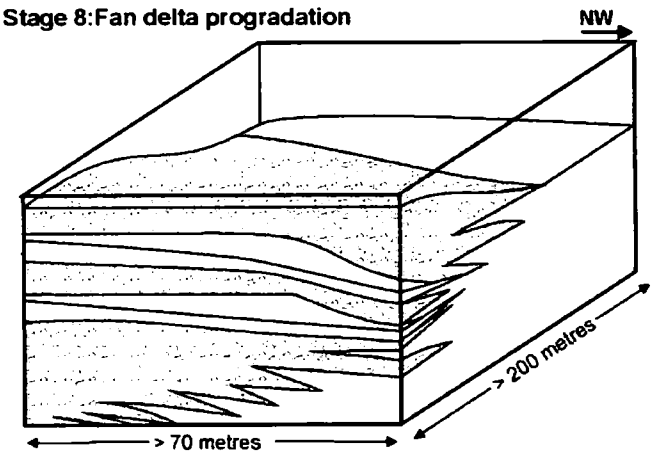
Figure 5.25 (cont.)

Stage 7: Siliciclastic hiatus: coral development



- carbonate develops as a clinoform, with the thickness of the unit increasing downslope
- upper surface is characterized by abundant columnar stromatolites, indicating a significant change to conditions detrimental for coral growth
- possible exposure and erosion of carbonate

Stage 8:Fan delta progradation



- fan-delta progradation towards the west
- gradual onlap and burial of carbonate

Facies

- Calcareous lithoclastic conglomerate
- Interbedded siliciclastic bioclastic packstone and calcareous litharenite
- Laminated coral platestone and siliciclastic bioclastic packstone (hardground interval)
- Coral mixstone
- Silty marl
- Laminar stromatolite

Figure 5.25 (cont.)

6. The effects of siliciclastic input on biota

6.1 Introduction

Carbonate production can occur in sites of active siliciclastic input e.g. the Red Sea (Purser 1987). There are numerous examples of carbonate provinces whose evolution has been directly influenced by siliciclastic sediment input e.g. the Great Barrier Reef (Larcombe and Woolfe 1999 a,b; Woolfe and Larcombe 1998 a,b), Kenyan coast (McClanahan and Obdura 1997) and Thailand (Tudhope and Scoffin 1994). Clastic material, as either bedload or suspended in the water column, has long thought to be detrimental to carbonate-producing marine organisms. Sediment may affect organisms in a variety of ways including physical burial and smothering, reduction in light levels through increased particulate matter suspended in the water column and changes in seawater chemistry.

A number of descriptive studies have been undertaken in order to understand the effects of siliciclastic input on biota in both modern (Tudhope and Scoffin 1994, Woolfe and Larcombe 1998a, 1998b, Larcombe and Woolfe 1999) and ancient depositional systems (Kumar and Saraswati 1997). Despite these numerous studies, there is limited published data quantifying the effects of clastic input on carbonate-producing biota. Studies by Lokier (2000), Lokier and Wilson (2002) and Wilson (*submitted*) aimed to quantify the effects of volcanoclastic and siliciclastic sedimentation on biota. The aim of this chapter is to describe and analyse quantitatively the response of carbonate organisms to siliciclastic input in the Calders study area. **Sections 6.1.1** to **6.1.3** aim to summarise the main effects of clastic input on carbonate-producing organisms, in particular photoautotrophs such as corals and larger benthic foraminifera. The methodology and results of acid digestion of samples from the Calders reefs are presented in **Sections 6.2** and **6.3**. **Sections 6.4** and **6.5** describe and discuss the effects of siliciclastic input on the benthic community in the Calders study area and relevant examples in the available literature. Samples from the Altorreal study area were unsuitable for acid digestion as coral-dominated carbonate units contain very little siliciclastic material. Qualitative observations regarding the relationship between biota and siliciclastics at Altorreal will be discussed where relevant. For a review of the depositional environments and the interaction between carbonate and siliciclastic facies in the area, refer to **Chapter 5**.

6.1.1 Physical burial and smothering of organisms

The effect of burial on a benthic marine organism will depend upon a number of factors including the rate and magnitude of sedimentation and the ability of an organism to excavate itself once buried. Burial is particularly hazardous to sessile benthic organisms such as corals, bryozoa and coralline algae that cannot excavate themselves if buried. An increase in amounts of sedimentation can often be related a decrease in coral growth rates (Tomascik *et al.* 1993).

The build-up of sediments on coral colonies can have lethal and sublethal effects (Gleason 1998). Sediment build-up can kill tissues through smothering (Rogers 1983, Stafford-Smith and Ormond 1992). Smothering leads to mortality through coral bleaching, tissue necrosis and/or anoxia (Riegl 1995, Riegl and Bloomer 1995, Wesseling 1999). On a more moderate scale, settling of sediments onto live coral surfaces can lead to significant increases in coral respiration while simultaneously decreasing net photosynthesis of endosymbiotic algae (Riegl 1995, Riegl and Branch 1997). Once coral polyps start to die, surrounding polyps are more suscepitable to being overwhelmed by sediment (Scoffin 1997).

Corals have several mechanisms for clearing sediments from their tissues. These include distension of tissues through the uptake of water, ciliary and tentacular action, and the secretion of mucus that carries away sediment (Roy and Smith 1971, Stafford-Smith and Ormond 1992). Surface cleaning using these mechanisms may prove detrimental to coral health as time and energy expended in shedding sediments may further decrease colony fitness by limiting resources available for other processes such as food capture, growth, tissue repair and reproduction (Hubbard and Pocock 1972, Kendall *et al.* 1975, Rogers 1990, Barnes and Lough 1999).

The benthic community structure in shallow marine settings largely reflects the nature of the substrate. Benthic communities within soft-bottomed, unstable sedimentary environments will be dominated by organisms with some degree of mobility thus are less susceptible to burial. Hard-bottom communities, where sedimentation rates are lower, often contain abundant sessile organisms as the need to excavate ones self does not arise (Heikoop *et al.* 1996). However, even mobile organisms are subject to burial if the amount of sediment input is very large. Certain coral species show a morphological adaption to sediment shedding, aiding their

survival in high settings with sedimentation rates. The development of funnel-shaped *Acropora clathrata* and *Tubinaria peltata* was observed in a high sedimentation environment off the coast of Natal, South Africa (Riegl *et al.* 1996). The funnel-shaped growth serves two different purposes. In calm waters with minor currents (corresponding to water depths of 18 to 25 metres) funnel shaped colonies act as sediment traps directing sediment to the apex of the funnel where coral tissue underwent necrosis, although tissues around the margins remained unaffected by sediment and healthy (Riegl *et al.* 1996). In relatively high-energy areas (corresponding to water depths of 8 to 14 metres) funnel-shaped corals were self-cleaning through a process of vortex-shedding (Riegl *et al.* 1996). Flume experiments have demonstrated that flow separation occurs on the funnel upstream edge as currents accelerate over the top (**Figure 6.1**). Turbulent eddies develop inside the funnel, capturing and transporting sediment away along the downstream margin (**Figure 6.1**). Current speeds as low as 30 to 90 cms-1 were enough to completely clean surfaces of sediment (Riegl *et al.* 1996).

The susceptability of corals to mortality from burial is largely species specific. Experiments have demonstrated that *Porites* can survive and recover from complete but short-term burial (up to 6 hours), although high rates of tissue damage and bleaching are recorded at burial times exceeding 20 hours (Wesseling *et al.* 1999). In Misima Island, Papua New Guinea, increased sedimentation rates associated with the construction of an open-cut gold mine resulted in high mortality rates of the Recent coral *Porites* (Barnes and Lough 1999). In cases where burial was incomplete, corals could tolerate up to 156 mg/cm2/d over a three year period (Barnes and Lough 1993). *Monastrea, Siderastrea* and *Colpophyllia* are tolerant of high sediment loading in addition to *Porites* (Edinger and Risk 1994). In comparison, *Acropora* is particularly susceptible to the effects of sedimentation, with complete colony mortality occuring after less than 6 hours burial (Wesseling *et al.* 1999). Burial mortality rates will be strongly affected by the sediment grainsize, with clays and silts more hazardous than sands (Wesseling *et al.* 1999, Thompson 1980).

Analysis of shallow coral communities in Kenya clearly demonstrates the relationship between coral generic abundances and sedimentation rates (McClanahan and Obdura 1997). Areas of low sedimentation are dominated by *Favia, Montipora*

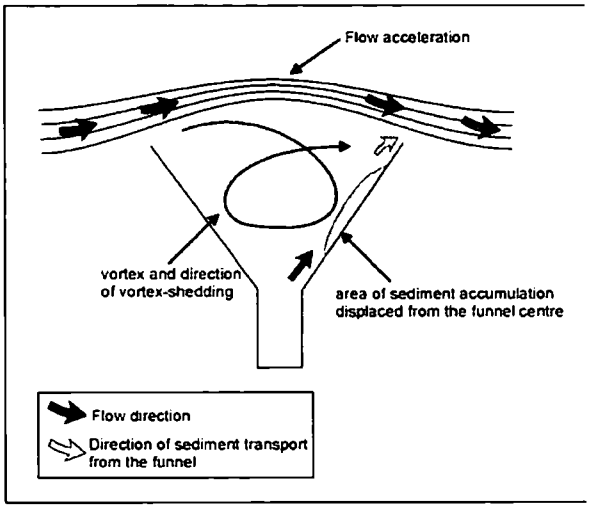


Figure 6.1 The function of funnel-shaped coral morphology in expelling sediment. Flow separation on the lip of the funnel generates turbulent eddies that expel collected sediment from the funnel center. After Rogers (1990).

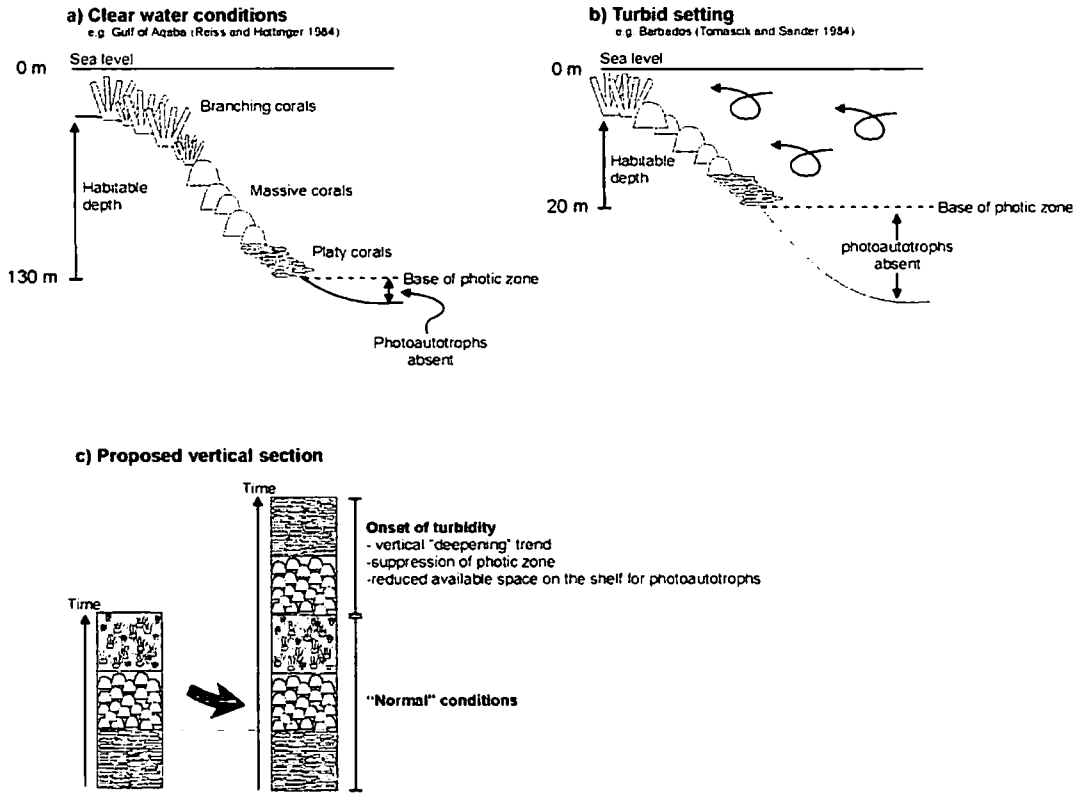


Figure 6.2 A comparison of coral morphology zonation under normal clear-water conditions (a) and turbid water conditions (b). Turbidity will suppress the depth of the photic zone, resulting in the movement of deeper-water morphologies into shallow water. In the rock record, the onset of turbidity may manifest itself as an apparent deepening-upward succession (c).

and *Poccillopora*, high sedimentation areas are dominated by *Echinopora, Galaxea, Hydnophora, Millepora* and *Platygyra*, and 'intermediate' areas are dominated by *Acropora, Astreopora, Favites* and *Porites* (McClanahan and Obdura 1997). Importantly, the abundance and generic richness does not vary considerably with regards to sedimentation, and thus it cannot be assumed that a decrease in coral abundance will accompany increased sedimentation rates (McClanahan and Obdura 1997).

6.1.2 Increasing turbidity

If clastic input is very fine grained, it can remain in suspension in the water column, attenuating available light levels. Turbidity has the greatest impact on benthic organisms that require light to produce energy via photosynthesis. Photosynthesis takes place using either chloroplasts (e.g. coralline algae) or through the incorporation of algal symbionts (e.g. corals, larger benthic foraminifera). In addition, precipitation of calcium carbonate in corals is dependent upon photosynthetic rates of endosymbiotic algae (Goreau 1959), thus reduced light intensity can lead to reduced rates of colony growth and algae endosymbiont productivity (Kendall *et al.* 1975, Rogers 1990). Many photoautotrophic benthic organisms are particularly at risk from suspended particulate material as they are unable to relocate themselves.

The amount of material in suspension, and the duration particles remain suspended, is fundamentally controlled by local and regional hydrodynamic conditions. In moderate to high-energy regimes, material may be carried into suspension, remaining in suspension for longer than in lower energy areas, thus having a greater impact. In such settings, clays and fine silts are often subject to resuspension, inducing turbidity even when sediment input from an external source is relatively low (Larcombe and Woolfe 1999). Patch reefs situated within the shallow inner shelf of the central Great Barrier Reef are often affected by frequent but short-lived turbidity events generated through a combination of wave and tidal activity and longshore currents (Larcombe *et al.* 2001). Clay and fine silts are largely derived *in situ* from a Holocene post-glacial muddy terrigenous wedge (Larcombe *et al.* 2001, Woolfe and Larcombe 1998).

The depth-related zonation of corals is often compressed in high-turbidity settings, with typically deep-water corals inhabiting relatively shallow habitats within

the photic zone (**Figure 6.2**). Corals that grow in turbid waters often show morphological adaptations to low light levels akin to those of a depth increase (Titlyanov and Latypov 1991, Wilson and Lokier 2002). Corals demonstrating platy, encrusting and corymbose morphologies, more typical of deeper water settings with low light intensities, occur in shallow waters (Titlyanov and Latypov 1991). Studies of turbid water reef development in the Miocene proto-Mahakam Delta, Borneo, demonstrate that extremely thin, platey coral morphologies are most common in low-energy environments that were characterised by high-suspended particulate matter (Wilson and Lokier 2002, Wilson, *submitted*). In addition, *Tubastrea*, an azooxanthellate that does not require sunlight, commonly occupies environmental niches in the patch reefs characterised by high turbidity. *Tubastrea* is present in a number of turbid marine settings in Indonesia (Tomascik 1997).

6.1.3 Changes in water chemistry

Coral and larger benthic foraminiferal communities typically develop in nutrient-deficient, oligotrophic marine waters (Hallock and Schlager 1986, Brasier 1995, Kumar and Saraswati 1997, Geel 2000, Hallock 2001). An input of siliciclastic material is often associated with rapid changes in seawater chemistry, in particular nutrient levels but also salinity and pH. Organisms with narrow tolerances to variations in salinity (stenohaline organisms) may die during such fluctuations (Lees 1975).

High nutrient levels may result from a terrestrial input, as sediment and/or run off within which organic material may be present as either organic plant material, or nitrate or phosphate ions in solution. High-nutrient environments are fundamentally unfavourable for carbonate production (Schlager 1992) since nutrients, particularly inorganic compounds, inhibit calcification rates (Simkis 1964). However, there is evidence to suggest that slightly elevated nutrient concentrations can result in increased growth rates in corals (Tomascik *et al.* 1997) and bivalves (Belda *et al.* 1993). This increased growth is accompanied by a decrease in skeletal density (Tomasacik *et al.* 1997).

A sudden increase in nutrients can have a number of detrimental effects on the benthic community. In high-nutrient settings, carbonate producers are effectively out-paced by non-skeletal competitors such as fleshy algae, soft corals and sponges

(Schlager 1992). Benthic community shifts from coral-dominated to coral/algae-dominated to macroalgae/sponge-dominated are primarily controlled by increased nutrient supply (Littler *et al.* 1991, Hallock 2001). An increase in the amount of fleshy algae, stimulated by high nutrient levels, will reduce the amount of stable substrate available for colonisation by other organisms (Hallock and Schlager 1986, Schlager 1992, Brasier 1995). Algae may also encrust and kill living corals, such as is observed off Belize and Florida (Antonius and Ballesteros 1998). Destruction of reef framework through bioerosion also increases with increasing nutrient levels as a consequence of an increased abundance of infaunal suspension feeders such as molluscs (Hallock and Schlager 1986, Schlager 1992, Edinger and Risk 1994).

Care needs to be taken when interpreting biota response to physiochemical parameters in the rock record. Initially, a change from coral-dominated to algal-dominated reefs in the Carribbean was related to pollution (and enhanced nutrient levels) (Bonem and Moses 1998). However, water sampling and further biota studies demonstrated that the increase in algal cover was actually due to reduced herbivory and mortality of the echinoderm *Diadema antillarium* (that fed exclusively on algae).

In summary, an input of terrigenous sediment to the shallow marine environment can have a number of detrimental effects on the benthic community. Organisms may be buried if the sediment input is large and they are unable to extract themselves. Changes in salinity may prohibit precipitation of carbonate. An increase in turbidity associated with high suspended sediment concentrations may reduce the habitable depth of the photic zone for organisms dependent on light. Sedimentation may result in organism mortality or alterations in body morphology. An increase in nutrients associated with sediment input can result in a switch from a benthic community dominated by corals and larger benthic foraminifera to a dominance of coralline and fleshy algae (depending upon the degree of herbivory).

6.2 Methodology

Sections at Calders were chosen for acid digestion since these sediments contained *in situ* carbonate-siliciclastic sediments (*sensu* Mount 1984). Bioclasts identified in the facies from the Calders study area ranged from less than 1 mm (i.e. smaller foraminifera) to nearly a metre (corals). In order to compare the relative abundance of different biota it has been necessary to combine visual abundance

estimates from outcrops and hand specimens with point counting data from thin sections. This method is therefore prone to a small degree of error.

Six distinct carbonate intervals intercalated with siliciclastics sediments characterise the Calders sedimentary succession (**Chapter 3, Figure 3.9**). In summary, the carbonate intervals represent colonisation of a siliciclastic substrate by corals, coralline and larger benthic foraminifera during times of significantly reduced siliciclastic input. The switch between siliciclastic and carbonate-dominated deposition is particularly well exposured along the B-124 road, represented by logs CA-4 (28 samples) and CA-7 (32 samples). Further less well-exposed areas of the Calders succession have also been chosen for acid digestion analysis as they can be directly correlated with the carbonate intervals of the road exposures. Further samples were selected from logs CA-2 (8 samples), CA-3 (14 samples), CA-5 (4 samples), CA-8 (5 samples), CA-9 (6 samples), CA-10 (4 samples) and CA-11 (4 samples). Samples analysed cover all of the facies identified in the Calders area.

In order to study the relationship between clastic input and relative abundance of biota, 106 whole rock samples of both carbonate and siliciclastic lithologies from the Calders area were treated with hydrochloric acid to determine the relative weight percentage of carbonate to non-carbonate material. The methodology for this procedure is explained in **Appendix 1**.

6.3 Relationship between biota and siliciclastic sediments in the study area

As with any with sedimentary study that involves reconstructing palaeoenvironments, it is not possible to establish every environmental parameter that influenced the development of the biota. For example, turbidity can only be inferred from the presence of clay-grade clastic material that must have been deposited from suspension. Turbidity itself is a transient hydrodynamic attribute of a sedimentary system that may leave no evidence in the rock record. It is possible, however, to observe the relative abundance and abundance trends of organisms and trace fossils with relation to amount of siliciclastic material, its grainsize and sedimentary structures. In this way, it is possible to hypothesise the characteristics of the depositional environment and how this has effected the development of the biota. In addition, care has been taken to differentiate between biota change as a consequence

of an environmental perturbation (such as increased sediment input or turbidity) or a natural ecological succession.

6.3.1 The relationship of biota to total non-carbonate content

The results of acid digestion of samples from the logged sections have been plotted against the weight percentage volume of carbonate within the lithologies examined in thin section. All data are illustrated in **Appendix 1**. Selected samples are presented in **Figure 6.3**. The data have also been plotted against the relative abundances of the main marine organisms for logged sections CA-2 (**Figure 6.4**), CA-3 (**Figure 6.5**), CA-4 (**Figure 6.6**) and CA-7 (**Figure 6.7**). Additional plots show the correlation between weight percentage of non-carbonate and corals (**Figure 6.8**) coralline algae (**Figure 6.9**), large free-living benthic foraminifera (**Figure 6.10**), *Gypsina* (**Figure 6.11**), larger encrusting foraminifera (**Figure 6.12**) and molluscs (**Figure 6.13**). In most cases, a broad relationship can be observed between relative abundance of biota and non-carbonate content.

6.3.1.1 Carbonate-siliciclastic interactions in the Calders succession

In the Calders study area, six carbonate intervals are situated within a predominantly siliciclastic succession. The measured thickness of the succession is approximately 130 m (**Figure 3.9**). Carbonate intervals range from < 2.5 m (interval 3) to 22.5 m (interval 6) in thickness.

The studied interval starts above the deep marine marly sandstones of the Vespella Formation, situated in the far south of the study area along the B124 road (**Figures 3.1** and **3.9**). Shallow marine sedimentation commenced with deposition of 2.5 m of fossilifierous sandstones, containing abundant *Nummulites* (**Figure 6.7**). *Nummulites* tests are up to 1.5 cm in diameter, and have intermediate lenticular morphologies. Down-dip, these nummulitic sands interdigitate with bioclastic rudstones and grainstones (**Figure 6.4**). The initiation of mixed carbonate-siliciclastic development is marked by the vertical transition of the nummulitic sandstones into siliciclastic-rich packstones, which contain abundant *Nummulites, Discocylina* and *Amphistegina* with coralline algae (as spheroidal concentric laminar and concentric columnar rhodoliths and detached laminar algal fragments), as well as other smaller

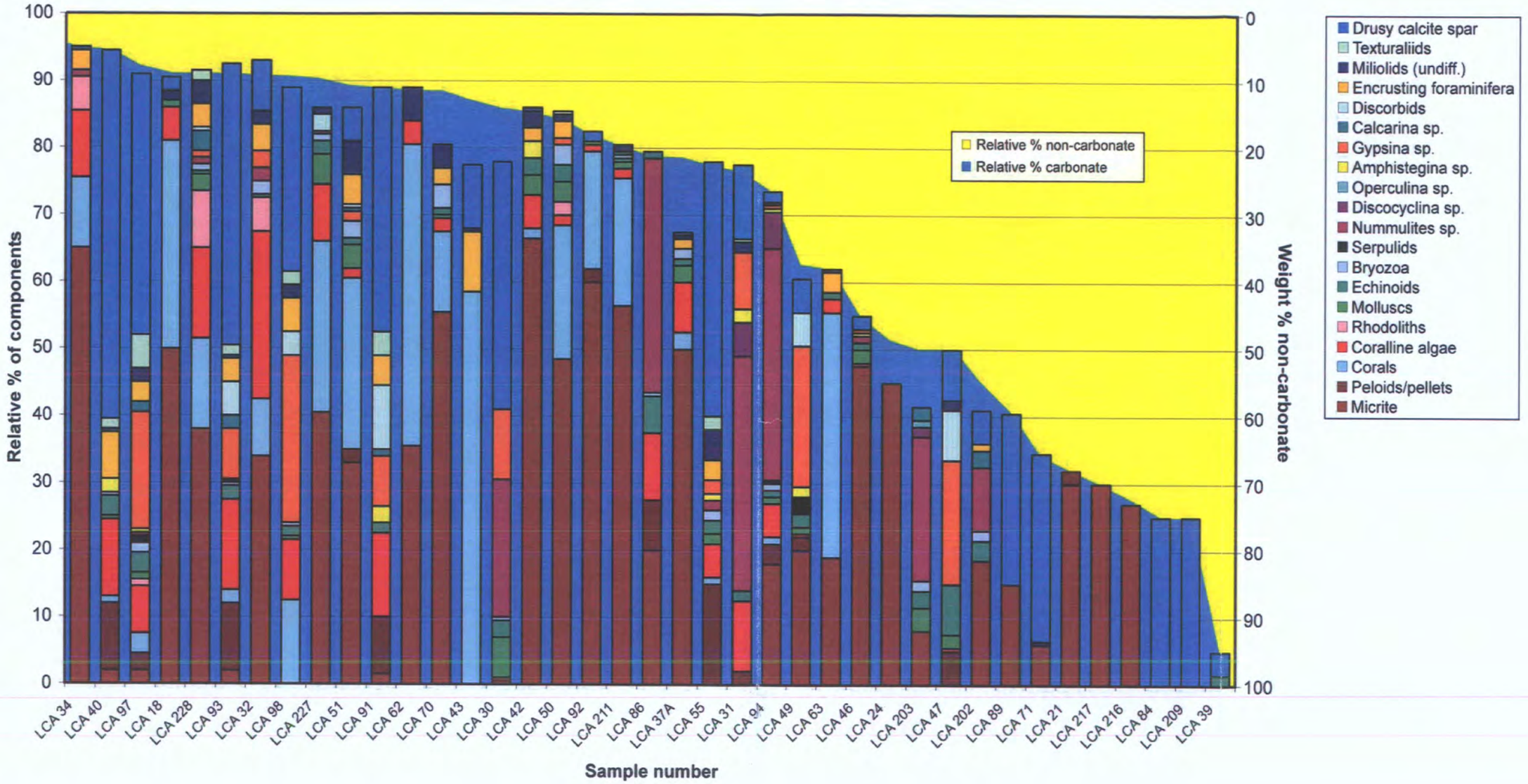


Figure 6.3 Graph to show the relationship between the relative abundance of biota and the weight percentage of non-carbonate material in selected samples from the Calders study area. Where the percentage of components does not reach 100%, the remainder is made up of a combination of non-carbonate material and very fine grained bioclastic debris.

benthic and encrusting foraminifera (**Figures 6.4** and **6.5**). The siliciclastic packstones contain, on average, 35-37 wt % of non-carbonate sediment. Bioturbation is present but rare, evident through alignment of bioclasts in large, domichnial burrows.

The initiation of carbonate-dominated deposition, and the development of a typical reef fauna, is represented by a sharp change in lithologies, from larger foraminifera and coralline algae dominated siliciclastic packstones to wackestones, packstones and rudstones (**Figures 6.5** and **6.7**). In outcrop, this change is associated with a colour change as beds go from brown and dark grey to pale grey and white. Wackestones, dominated by largely *in situ* delicate branching corals, often contain a high percentage (up to 26.5 wt. %) of clay to silt grade non-carbonate (**Figure 6.7**). Although preserved intact, corals do not make up a framework. Packstones and rudstones contain abundant corals that are often encrusted by coralline algae and foraminifera such as *Gypsina, Fabiania, Chapmanina, Haddonia* and victoriellids (**Figures 6.4**, **6.5** and **6.7**). Coral fragments act as the nucleus for tightly, concentric spheroidal foralgal rhodoliths.

The cessation of carbonate production is marked by deposition of a 3.0 m thickness of non-fossiliferous sandstones and conglomerates (**Figure 6.7**). The contact is sharp but non-erosional. These siliciclastic lithologies contain up to 25 % carbonate as intergranular, drusy calcite cement. The conglomerates contain reworked *Nummulites* tests, and pass vertically into nummulitic siliciclastic packstones (the second carbonate interval-**Figures 6.5** and **6.7**). The siliciclastic packstones contain 22 to 35 wt. % non-carbonate material, and are dominated by moderately reworked *Nummulites* and coralline algae, with *Amphistegina* and *Gypsina* (**Figure 6.7**). *Nummulites* test size increases vertically through the siliciclastic packstone beds (from < 1 cm to 3.5 cm), although larger tests are quite robust, giving an approximate diameter to thickness ratios of 0.07 to 0.2. As before, coralline algae occur as laminar, tightly concentric spheroidal rhodoliths with common laminar and warty algal fragments in the matrix. Rare coral fragments are identified in siliciclastic packstones, although corals are not observed *in situ*. The siliciclastic packstones are succeeded vertically by approximately 2.5 m of coral floatstones and bindstones that comprise the second carbonate interval (**Figures 6.5** and **6.7**). Corals initially show platy morphologies (**Figure 6.7**), although are rapidly supplanted by branching forms.

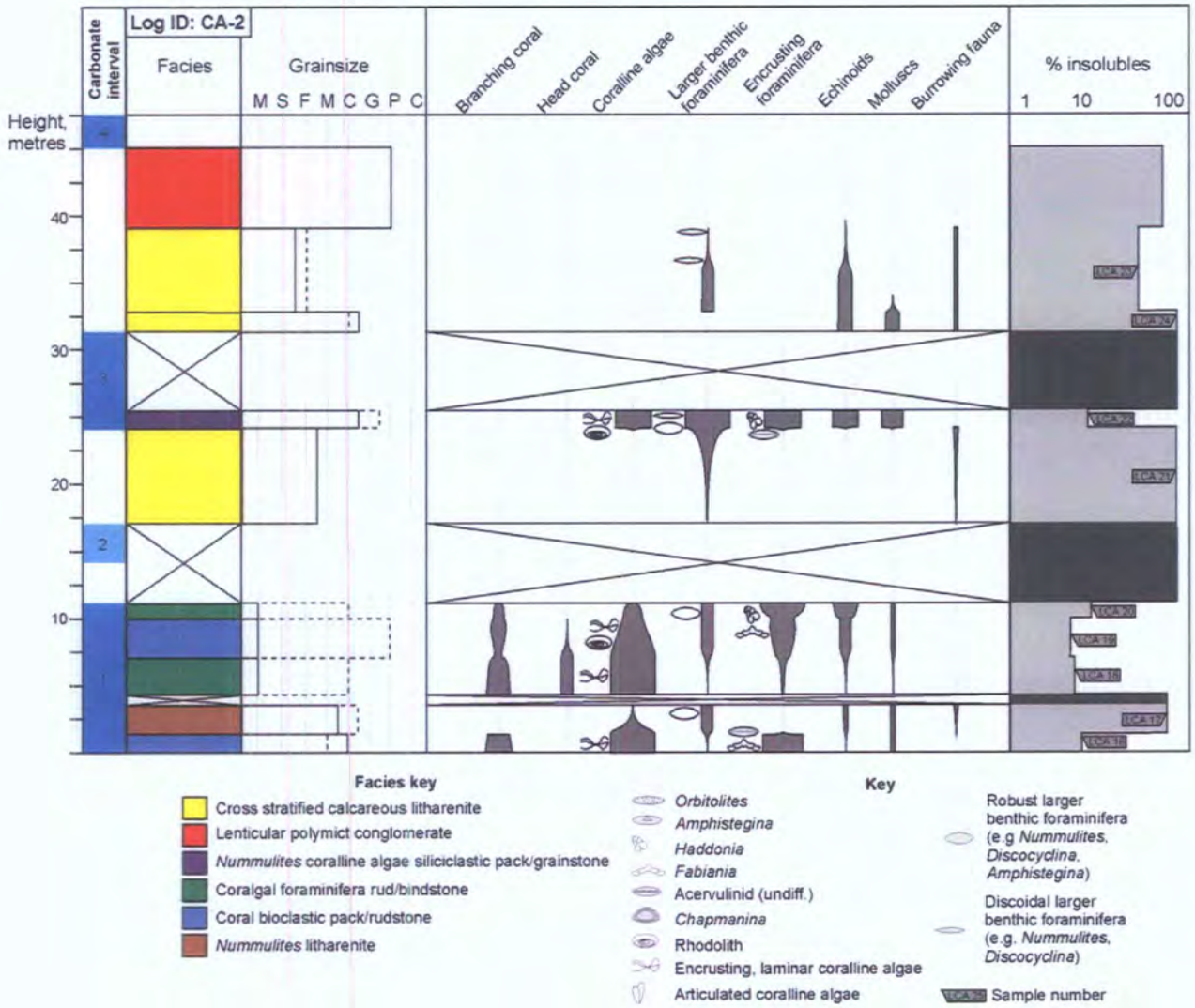


Figure 6.4 The total weight percentage of non-carbonate component compared with the relative abundances of the principal biota within facies of logged section CA-2. Abundances of organisms are based upon thin section point counting data in combination with hand specimen and outcrop observations.

Corals exhibit thin, laterally restricted coralline algae crusts that are often detached and preserved floating in the muddy matrix. Foraminfera are dominated by miliolids. Echinoids, molluscs and fenestrate bryozoa are also present in appreciable amounts in this carbonate interval. The micritic matrix contains up to 20.5 wt. % non-carbonate material interpreted as optically non-resolvable clays (**Figure 6.5**). The top of the second carbonate interval is marked by a *Nummulites*-dominated siliciclastic packstone, containing up to 26.5 wt. % non-carbonate (**Figures 6.5** and **6.7**). This lithology is not observed at the top of any of the other carbonate intervals and its significance will be discussed in **Section 6.4.1.2.1**.

The second carbonate interval is suceeded vertically by a further thickness (up to 10 m) of cross-stratified sandstones (**Figures 6.4, 6.5 and 6.7**). Some sandstone horizons are bioturbated and contain molluscan debris (**Figure 6.7**). Stratigraphically isolated wacke/packstone units, containing flattened discoidal *Nummulites* and *Discocylina,* occur within the sandstone (**Figure 6.7**). The sandstones culminate in a conglomerate that passes vertically into the third carbonate interval (**Figures 6.4, 6.5** and **6.7**). The third carbonate interval is relatively thin (< 2.5 m), and as observed in lower parts of the succession, is composed of a siliciclastic packstone dominated by *Nummulites*, that passes vertically into a siliciclastic *Nummulites, Discocylina* and rhodolith-dominated packstone (**Figure 6.7**). Rhodoliths are composed of coralline algae and encrusting foraminifera (*Gypsina, Fabiania* and *Haddonia*), and have laminar concentric spheroidal growth forms, with coral fragments as nucleii. The incorporation of foraminifera has often resulted in mis-shapen rhodoliths (discussed in **Section 6.4.1.4**). The siliciclastic packstones are succeeded vertically by rud/bindstones that are dominated by corals with multiple (up to 1 cm thick) laminar crusts of coralline algae and foraminifera (*Gypsina, Haddonia* and victoriellids). These carbonate lithologies contain < 10 % non-carbonate material (**Figure 6.5**). Larger benthic foraminifera such as *Nummulites* and *Discocylina,* are absent although small, robust forms of *Amphistegina* are moderately abundant (**Figure 6.5**).

The third carbonate unit is succeeded by a 10 m thickness of bioturbated siliciclastics (**Figure 6.7**). Body fossils are absent. Burrows, up to 10 cm in length, are oriented normal to bedding. The sandsones may comprise up to 45 wt. % carbonate as intergranular, drusy calcite cement. The fourth carbonate interval follows the same

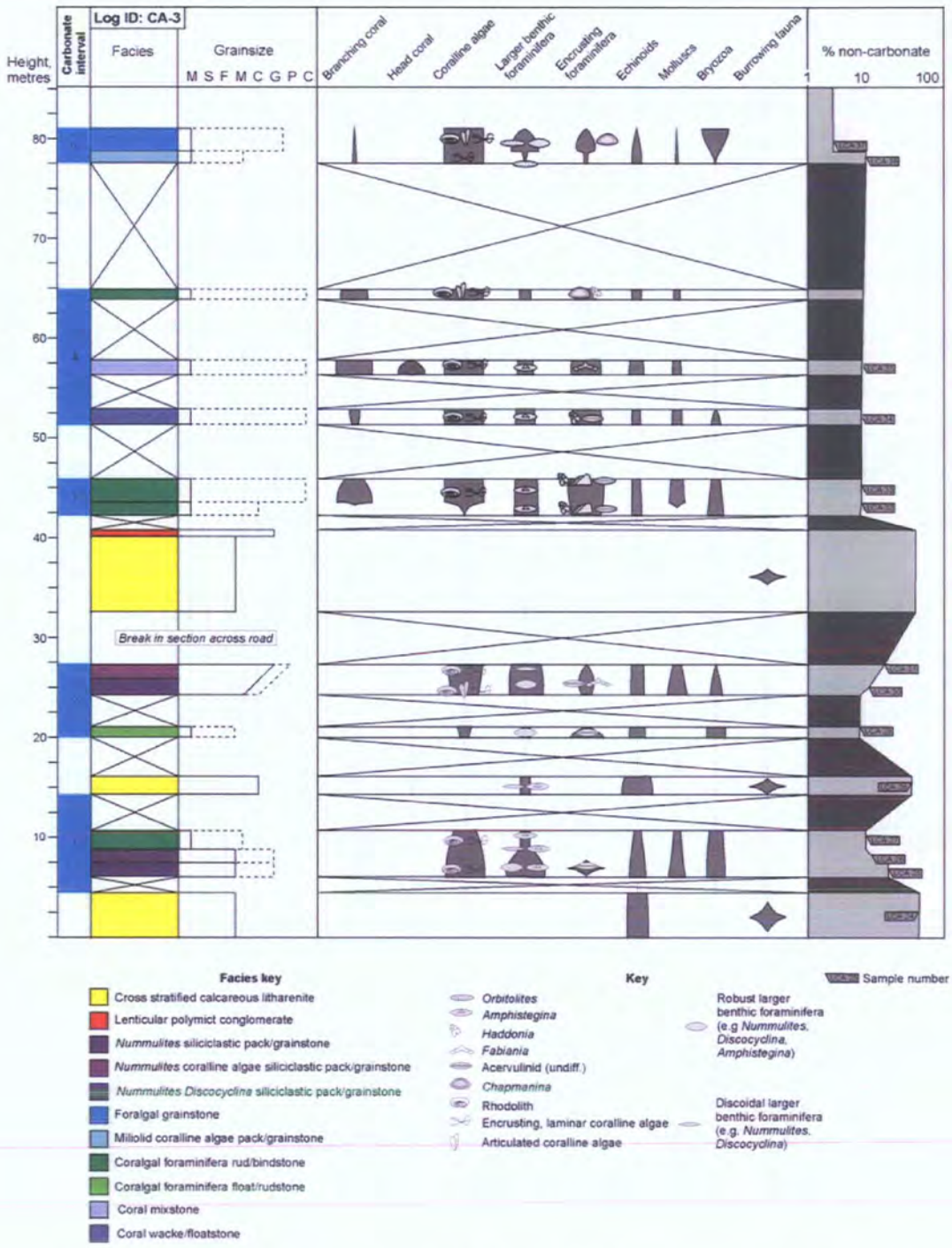


Figure 6.5 The total weight percentage of non-carbonate component compared with the relative abundances of the principal biota within facies of logged section CA-3. Abundances of organisms are based upon thin section point counting data in combination with hand specimen and outcrop observations.

pattern as stratigraphically lower intervals, with siliciclastic-rich *Nummulites* and *Discocylina* packstones passing vertically into 7 to 10 m of coral-dominated floatstones, rudstones, bindstones and mixstones containing < 10 wt. % non-carbonate material (**Figure 6.5** and **6.7**). Coralline algae and large foraminifera such as *Haddonia, Gypsina* and *Fabiania* often encrust corals. Crusts are up to 1 cm thick. The fourth carbonate unit is succeeded by a poorly-exposed sandstone unit whose thickness is around 7.5 m (**Figure 6.5**).

The fifth carbonate unit is approximately 22.5 m thick, and unlike observed for the underlying units, the larger foraminiferal, siliciclastic-rich packstone is absent (**Figure 6.6**). This carbonate interval is composed of coral, coralline algae and larger encrusting foraminifera dominated floatstones, rudstones, grainstones, rud/bindstones and mixstones (**Figure 6.6**). The total non-carbonate content is less than 20 wt. %, although a sharp-based, coarse siliciclastic unit is present towards the base of the fifth carbonate unit (**Figure 6.6**). Coralline algae occur as laminar to columnar spheroidal rhodoliths, and as thick concentric crusts on corals. Large encrusting foraminifera, in particular *Fabiania, Haddonia, Gypsina, Chapmanina* and victoriellids, are also observed encrusting corals, and as a constituent of foralgal rhodoliths.

The transition between the fifth carbonate unit and overlying siliciclastics is very well exposed. The fifth carbonate unit culminates in a coral mixstone dominated by branching corals and containing less than 10 wt. % non-carbonate grains (**Figure 6.6**). The coral mixstone is succeeded initially by a thin coral bioclastic pack/rudstone (0.45 m) containg 26 wt. % non-carbonate and a 0.4 m thick bioclastic siltstone containing 45 wt. % non-carbonate (**Figure 6.6**). Corals are absent in the siltstone, although large, discoidal *Nummulites* are preserved within burrows.

The fifth carbonate interval is followed by a 20 m thickness of siliciclastics, dominated by non-fossilifierous, bioturbated sandstones interbedded with mudstones, shales and siltstones. Rare large, flattened *Operculina* are identified within mudstones. *Operculina* tests are very thin, with a typical test thickness of 1 to 2 mm with approximate diameter to thickness ratios of 0.02 to 0.1. Isolated floatstone lithologies, containing 34 wt. % non-carbonate, are present within the siliciclastics (**Figure 6.6**). Floatsones are composed of delicate branching corals with rare thin coralline algae crusts.

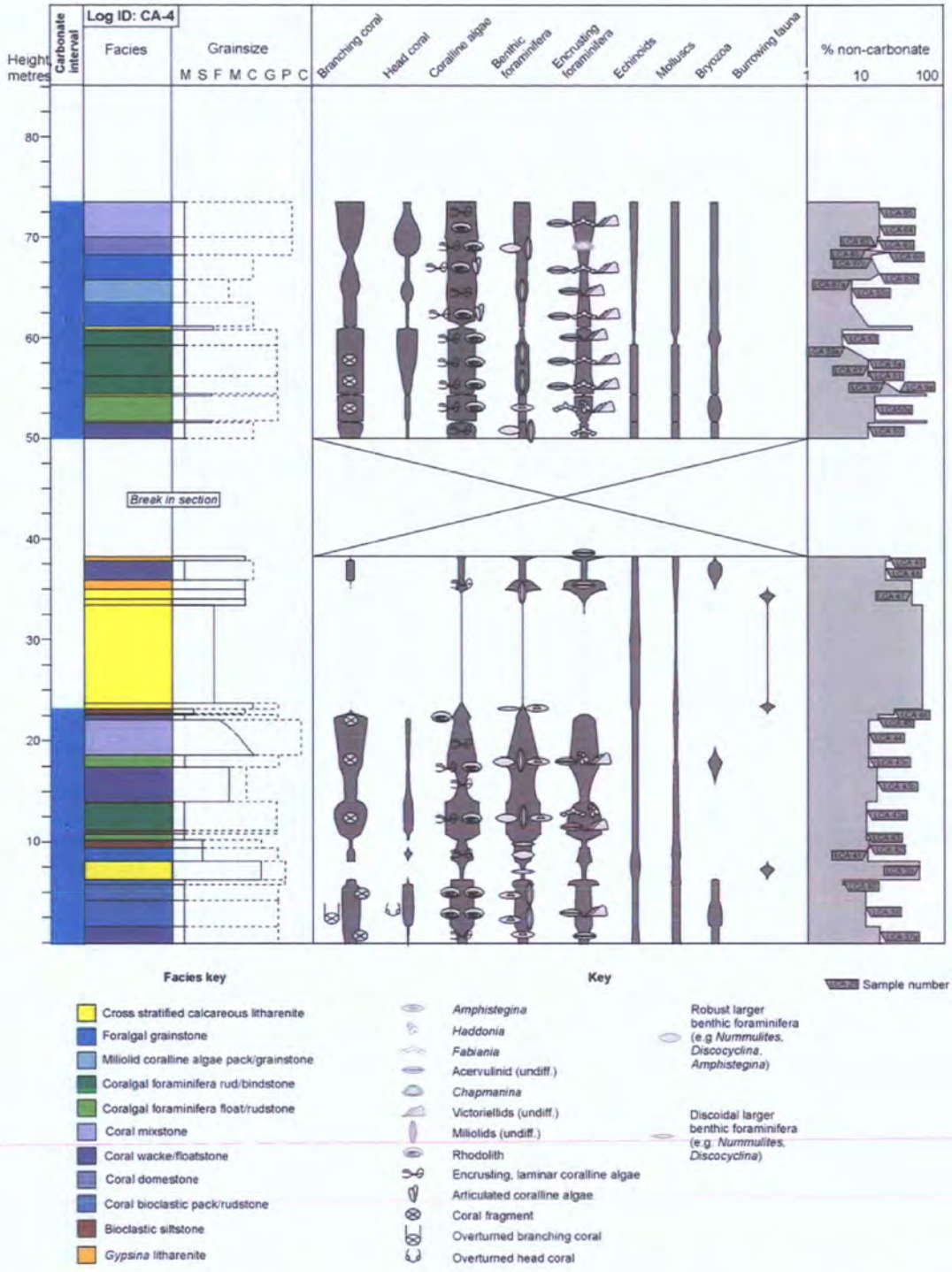


Figure 6.6 The total weight percentage of non-carbonate component compared with the relative abundances of the principal biota within facies of logged section CA-4 (parts a and b). Abundances of organisms are based upon thin section point counting data in combination with hand specimen and outcrop observations.

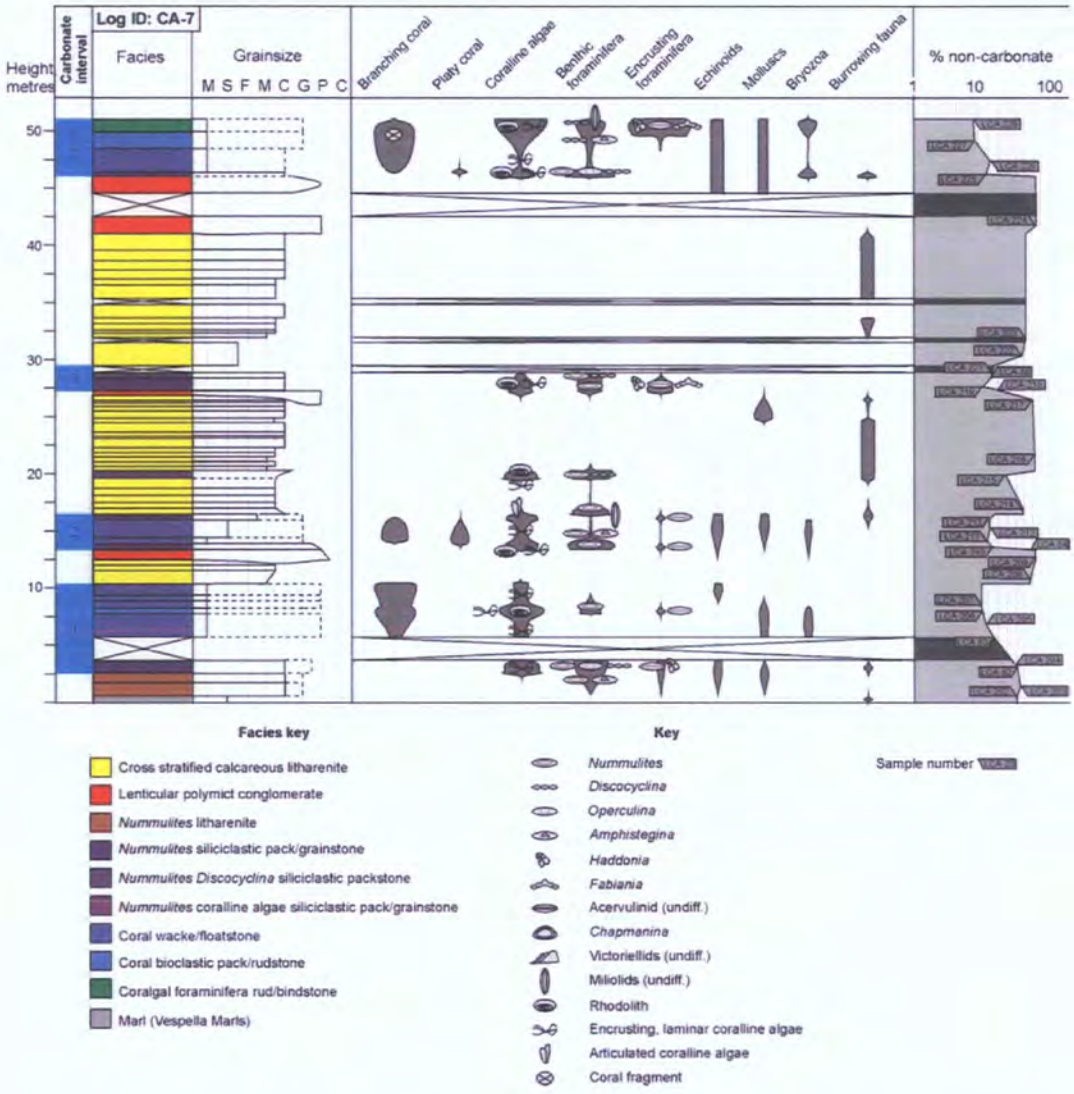


Figure 6.7 The total weight percentage of non-carbonate components compared with the relative abundances of the principal biota within facies of logged section CA-7 (parts a and b). Abundances of organisms are based upon thin section point counting data in combination with hand specimen and outcrop observations.

The sixth carbonate interval is approximately 22.5 m in thickness, and is dominated by corals, coralline algae and larger encrusting foraminifera (**Figure 6.6**). The lower 10 m is composed of mainly reworked facies with abundant rhodoliths and laminar coralline alage fragments. Rhodoliths, up to 7.5 cm in diameter, are often composed of coralline algae and foraminifera such as *Gypsina, Haddonia* and *Fabiania*. Rhodolith morphologies include spheroidal concentric laminar and spheroidal columnar and branching. Incorporation of foraminifera has often resulted in mis-shapen rhodoliths (**Section 6.4.2.2**). These lower lithologies contain 5.5 to 27 wt. % of fine-grained, non-carbonate material. The upper 5 metres of this carbonate interval is dominated by *in situ* corals. A head coral-dominated unit (2 m thickness) is succeeded by a coral mixstone with branching, head, massive and platy colony morphologies (**Figure 6.6**). These lithologies contain 22 to 38 wt. % fine-grained non-carbonate material. The contact of the final carbonate interval and the overlying siliciclastic sediments cannot be observed directly in the field, although a siliciclastic unit 4 m in thickness and containing *Gypsina*, stratigraphically separates the top of the Calders reef and the Sant Amanc succession (**Figure 3.62**).

6.3.1.2 The relationship of weight percentage non-carbonate sediment to biota

From the plots presented as **Figures 6.3** to **6.13,** the following relationships between biota and the amount of non-carbonate sediment can be observed:

- Branching corals are abundant in sediments containing up to 38.0 % clay to silt-grade non-carbonate material (**Figures 6.3** and **6.8**). Corals are not observed where the weight percent of non-carbonate material exceeds this value (**Figure 6.3**).
- Corals are most abundant in sediments containing around 10 wt. % of non-carbonate material (**Figure 6.8**).
- Coralline algae are observed in lithologies containing up to 38.5 wt. % non-carbonate material, although at higher percentages fragments show moderate abrasion and therefore have been reworked (**Figure 6.9**).
- Coralline algae are most abundant in sediments containing 10 to 30 wt. % non-carbonate sediment (**Figures 6.3 and 6.9**).

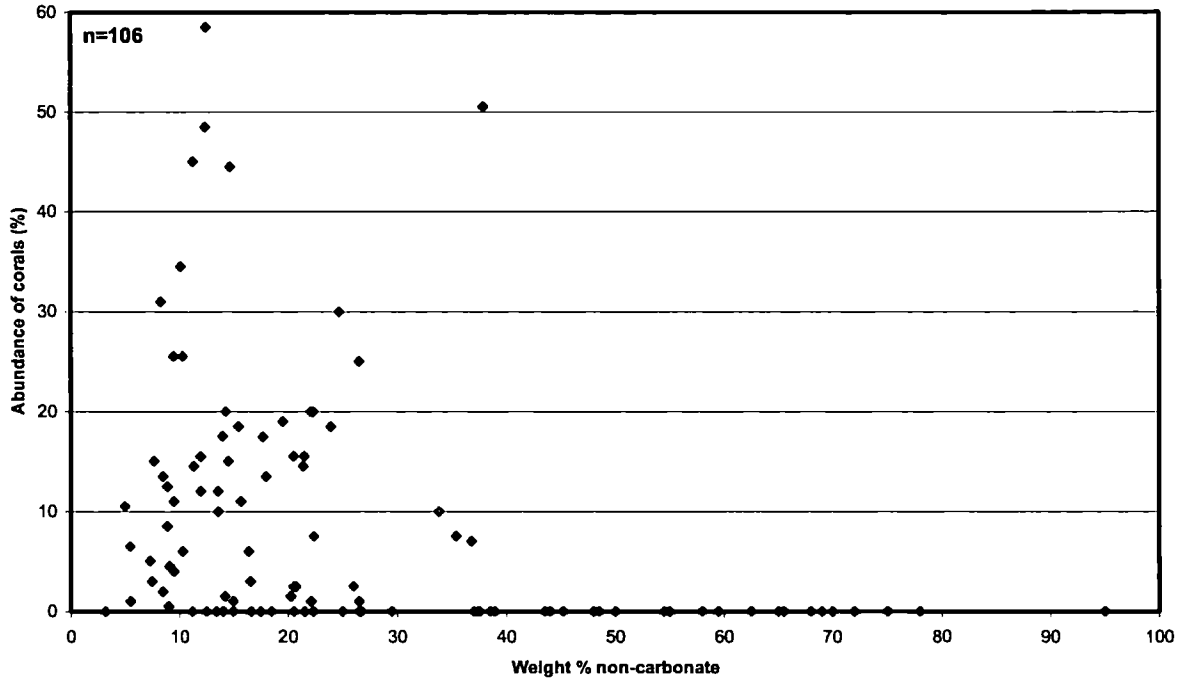


Figure 6.8 Plot of coral abundance (%) against whole-rock non-carbonate content for samples from logs CA-2 to CA-11 (106 samples). Coral abundances were determined through point counting.

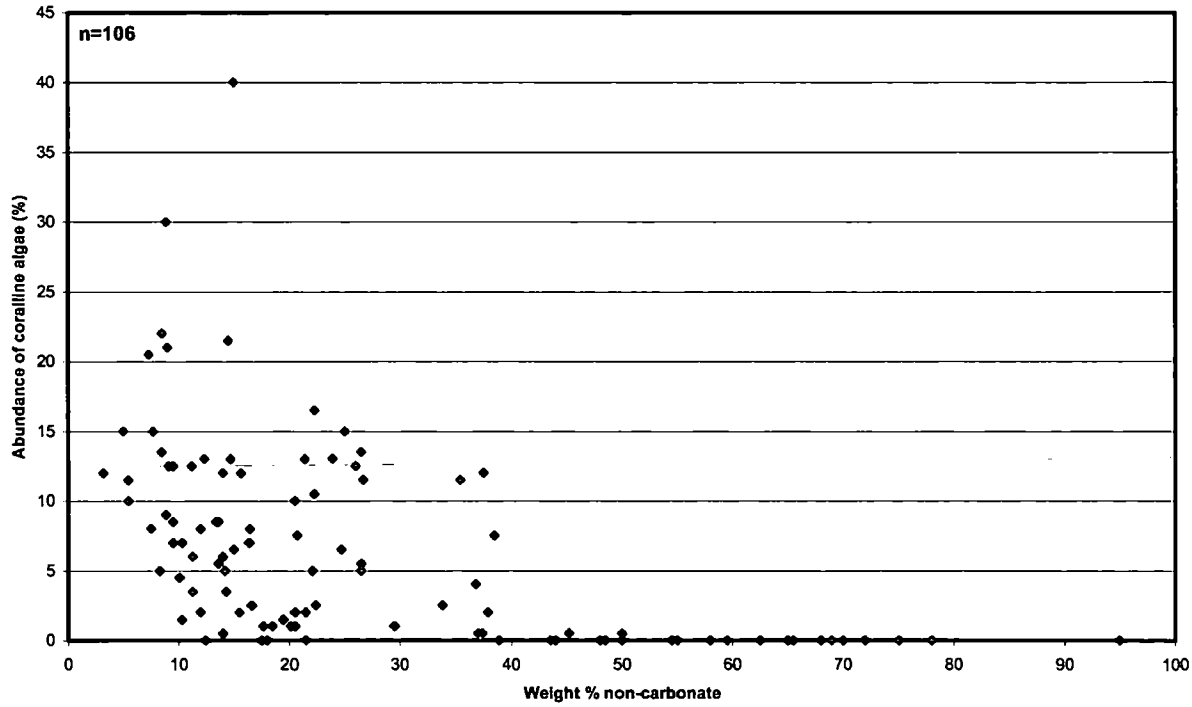


Figure 6.9 Plot of coralline algae abundance (%) against whole-rock non-carbonate content for samples from logs CA-2 to CA-11 (106 samples). Coralline algae abundances were determined through point counting.

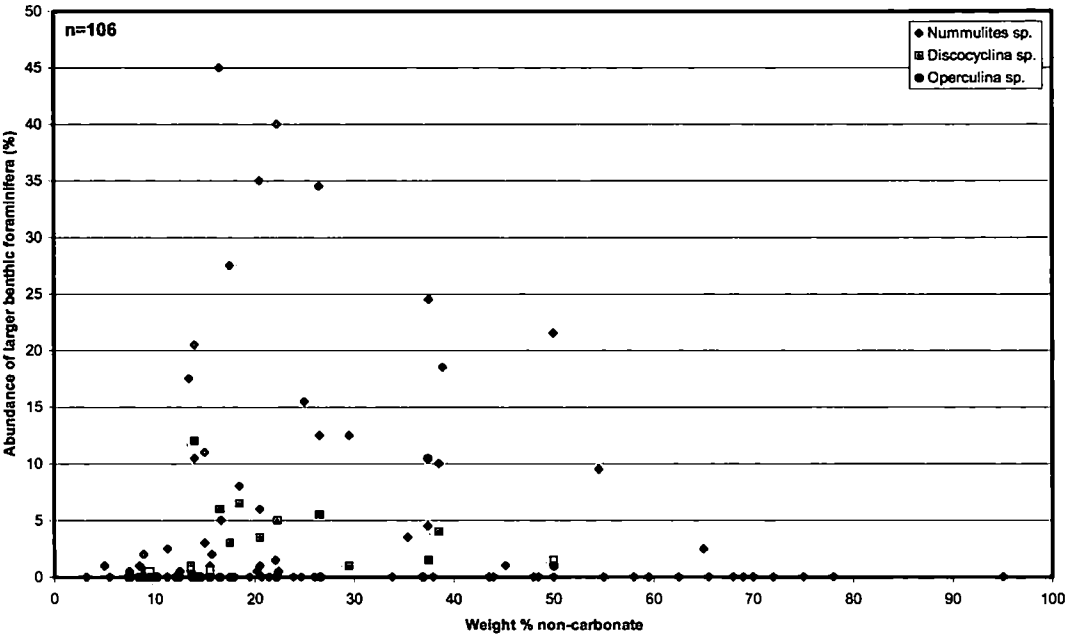


Figure 6.10 Graph of larger benthic foraminifera abundance within hand specimen samples against whole rock non-carbonate content for samples from logs CA-2 to CA-11 (106 samples). Foraminifera abundances were determined through point counting.

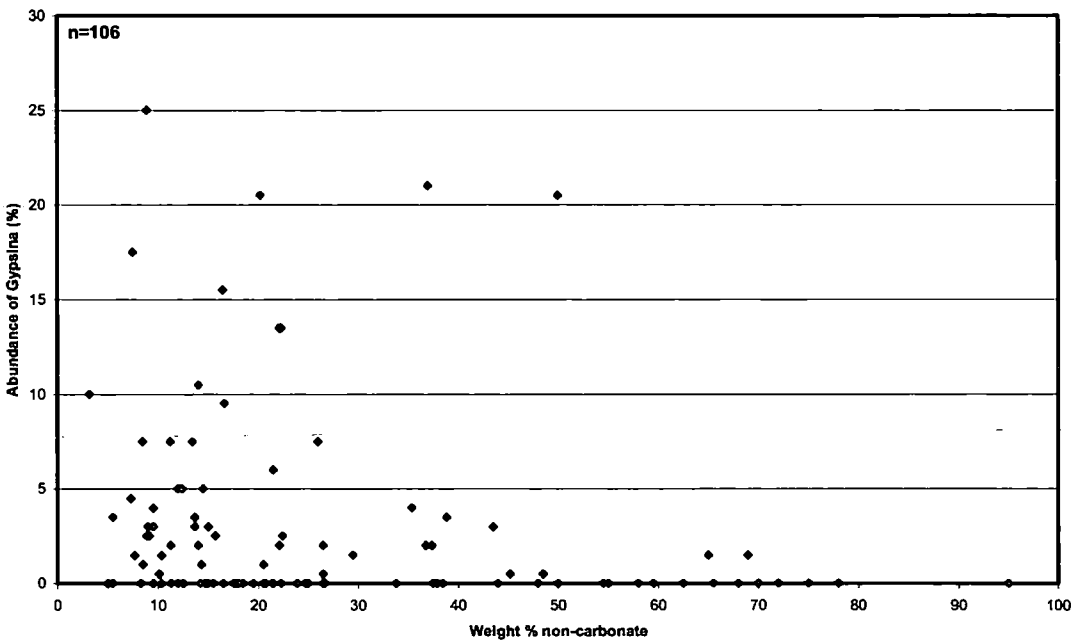


Figure 6.11 Graph of *Gypsina* abundance against whole rock non-carbonate content for samples from logs CA-2 to CA-11 (106 samples). *Gypsina* abundances were determined through point counting.

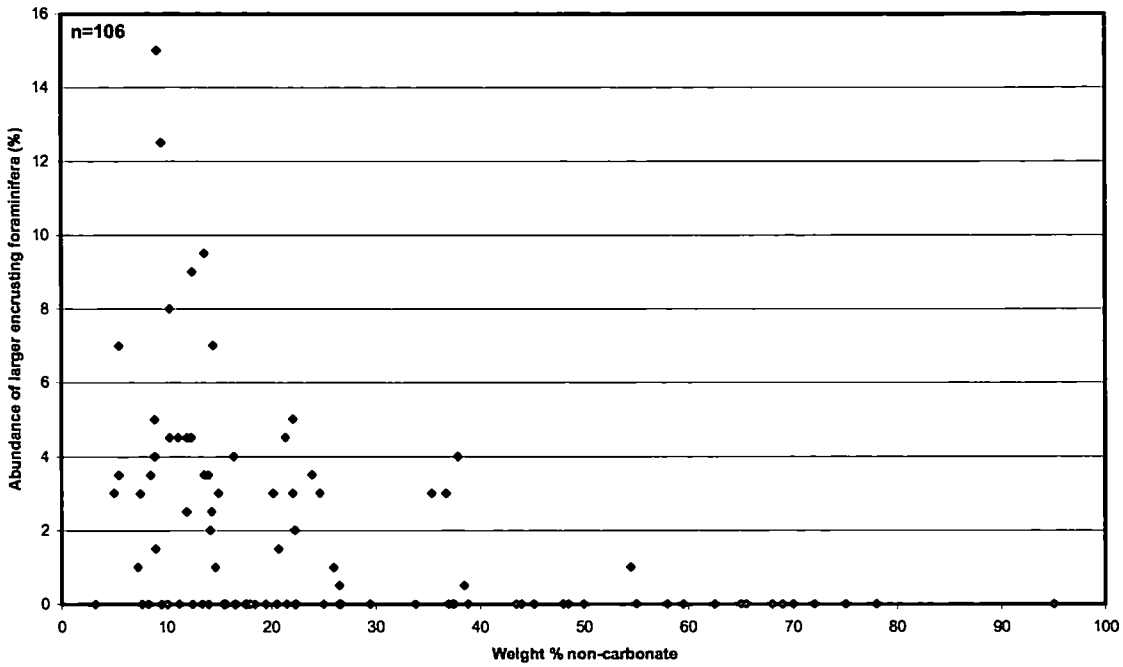


Figure 6.12 Graph of larger encrusting foraminifera abundance against whole rock non-carbonate content for samples from logs CA-2 to CA-11 (106 samples). Foraminifera abundances were determined through point counting.

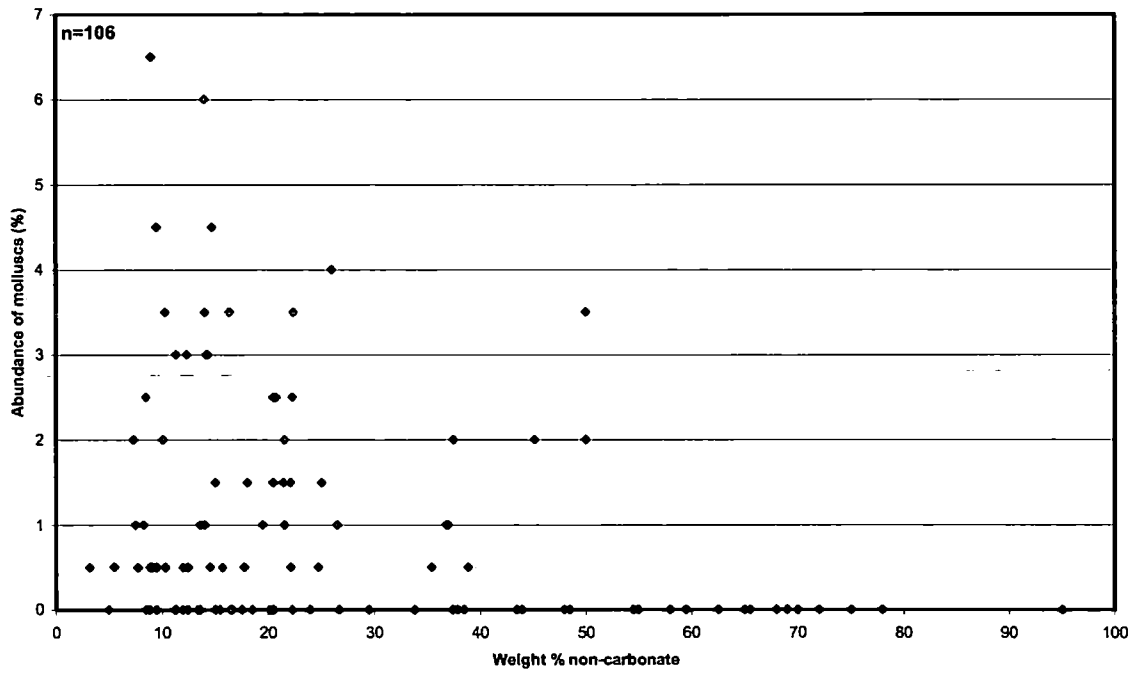


Figure 6.13 Graph of molluscan abundance against whole rock non-carbonate content for samples from logs CA-2 to CA-11 (106 samples). Mollusc abundances were determined through point counting.

- The larger benthic foraminifera *Nummulites, Discocyclina* and *Operculina* occur at the highest levels of non-carbonate sedimentation, and are moderately common in sediments containing up to 65 wt. % non-carbonate material (**Figure 6.10**).
- The encrusting foraminifera *Gypsina* is abundant in sediments containing 3.2 to 69 wt. % non-carbonate (**Figure 6.11**). *Gypsina* is most abundant in sediments containing 3 to 30 wt. % non-carbonate sediment (**Figures 6.3 and 6.11**).
- Larger encrusting foraminifera are abundant in sediments with a narrow range of weight percentage non-carbonate (**Figures 6.3 and 6.12**). Encrusting foraminifera are present in sediments containing up to 54.5 wt. % non-carbonate, although are more common are levels of <10 wt. % (**Figure 6.3**).
- Molluscs are able to tolerate a wide variety of levels of non-carbonate, and occur in moderate abundances in sediments ranging from <10 to 50 wt. % non-carbonate (**Figure 6.13**).

6.3.2 Relationship of biota to non-carbonate sediment grainsize

The median grain diameters of the largest non-carbonate grains within selected samples were calculated using the method described in **Appendix 1**. The relative abundance of carbonate components, determined through point counting, has been plotted against the maximum siliciclastic grainsize (**Figures 6.14** to **6.18**). Several relationships are observed:

- It has already been noted (**Section 6.3.1.1**) that branching corals are present in sediments containing large quantities of non-carbonate material (up to 38 wt. %). Branching corals are able to tolerate a variety of grainsizes of non-carbonate material, although corals are more abundance where the median grainsize is < 250 μm (and are particularly common where non-carbonate grainsize is 4 μm) (**Figure 6.15**).
- Branching corals are absent where the median clastic grainsize > 1000 μm (**Figure 6.15**).
- Coralline alage are present at non-carbonate grainsizes up to 4000 μm, although are most abundant where the grainsize ranges from 4 to 1000 μm (**Figure 6.16**).

- Large benthic foraminifera (*Nummulites* and *Discocyclina*) are abundant at grainsizes ranging from 4 to 1000 μm (**Figure 6.17**). *Operculina* is abundant where non-carbonate grainsize <125 μm.
- The encrusting form of *Gypsina* is able to tolerate a wide variety of non-carbonate grainsizes, although is most abundant where the grainsize is around 1000 μm (**Figure 6.18**).

6.4 Responses of biota to siliciclastic sediment input

Siliciclastic input has been a significant influence on the benthic marine biota in the Calders study area. Importantly, this input has not always proved to be detrimental to all organisms present. Corals, coralline algae and larger benthic foraminifera are the most important organisms present within the Calders succession, thus form the focus of the investigation of the effects of clastic input on biota. Other organisms considered include large encrusting foraminifera, molluscs and soft-bodied infaunal organisms.

6.4.1 Corals

Corals are one of the most important biotic components of the carbonate intervals of the Calders depositional succession, although corals are absent from siliciclastic lithologies. This is to be expected, considering the siliciclastic facies were deposited within moderate to high-energy settings characterised by unstable substrates (**Section 3.3.3** and **3.3.4**). Shifting sands would have effectively inhibited colonisation by corals in their larval stages.

The relative abundance of corals shows a correlation to both the non-carbonate sediment abundance (**Section 6.3.1.1**) and grainsize (**Section 6.3.2**). Corals are absent from lithologies containing > 38 wt. % non-carbonate sediment (**Section 6.3.1.1, Figure 6.12**), likely due to the effects of smothering (cf. Rogers 1983, Stafford-Smith and Ormond 1992). Corals are absent from lithologies where the non-carbonate grainsize exceeds 1000 μm (**Figures 6.14** and **6.15**). It is suggested that this is related to damage of coral polyps through ingestion of coarse angular grains (cf. James and Kendall 1992) and/or to the abrasion of soft tissue resulting in mortality (Woolfe and Larcombe 1999).

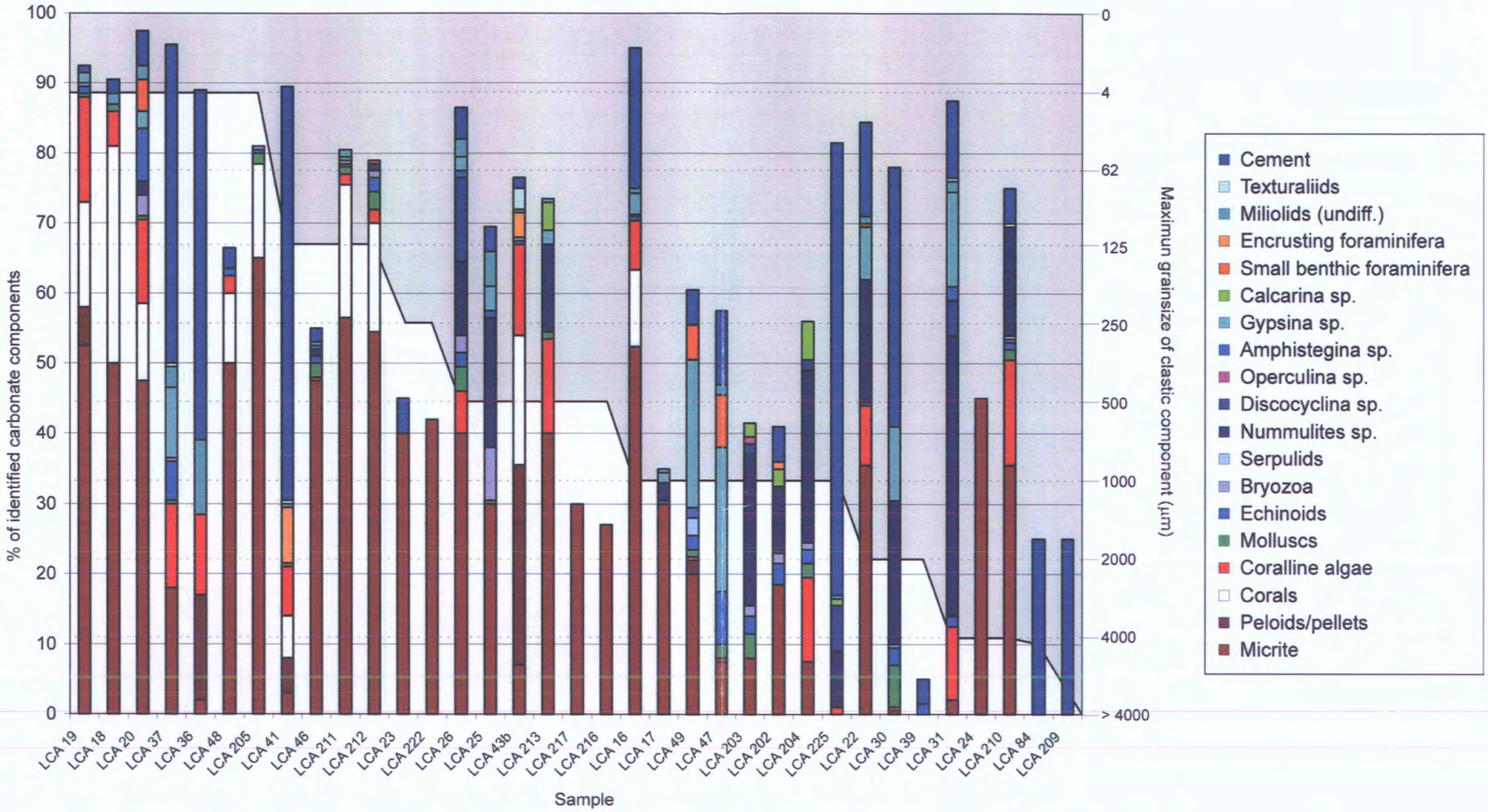


Figure 6.14 Graph showing the relationship between the principal carbonate components with the grainsize of the largest siliciclastic grain present. Where the sum of components does not reach 100 %, this is accounted for by non-carbonate material and unidentifiable bioclastic debris.

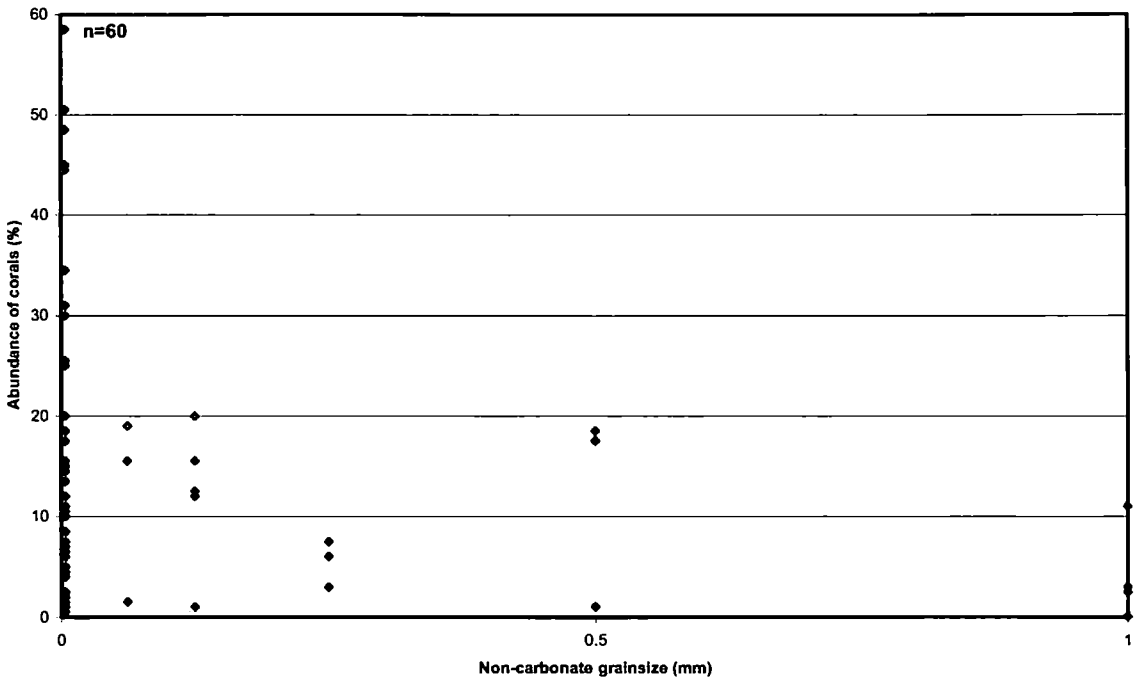


Figure 6.15 Plot of coral abundance (%) against median grain size of the largest non-carbonate clasts from selected samples in the Calders area (60 samples).

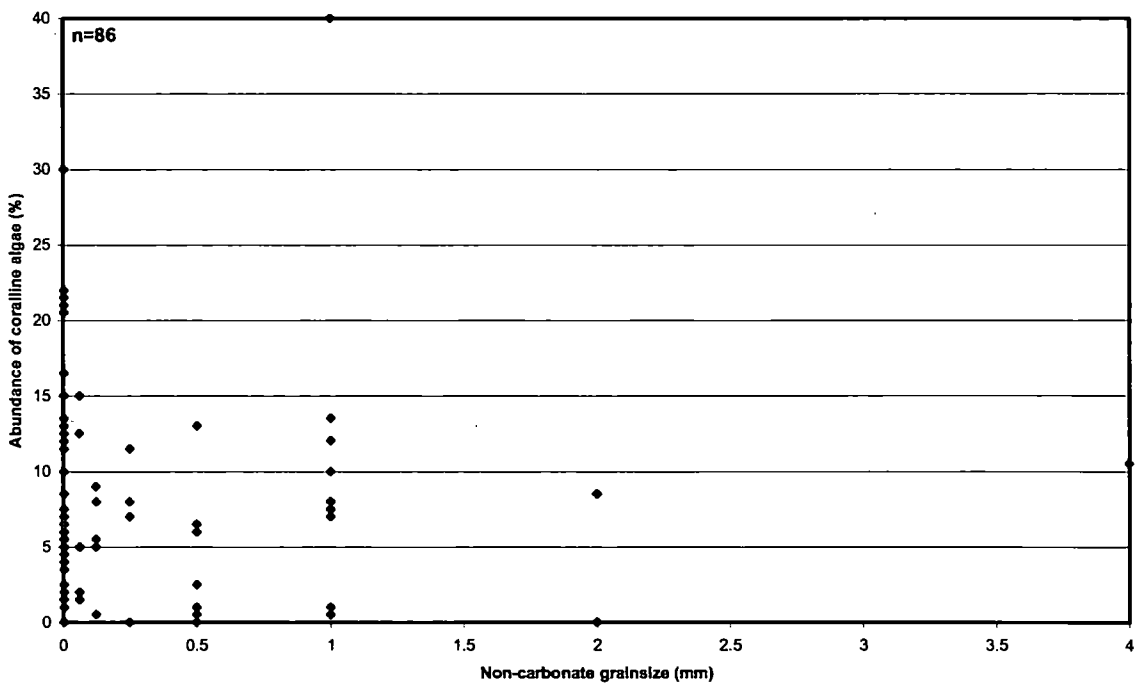


Figure 6.16 Plot of coralline algae abundance (%) against median grain size of the largest non-carbonate clasts from selected samples in the Calders area (86 samples).

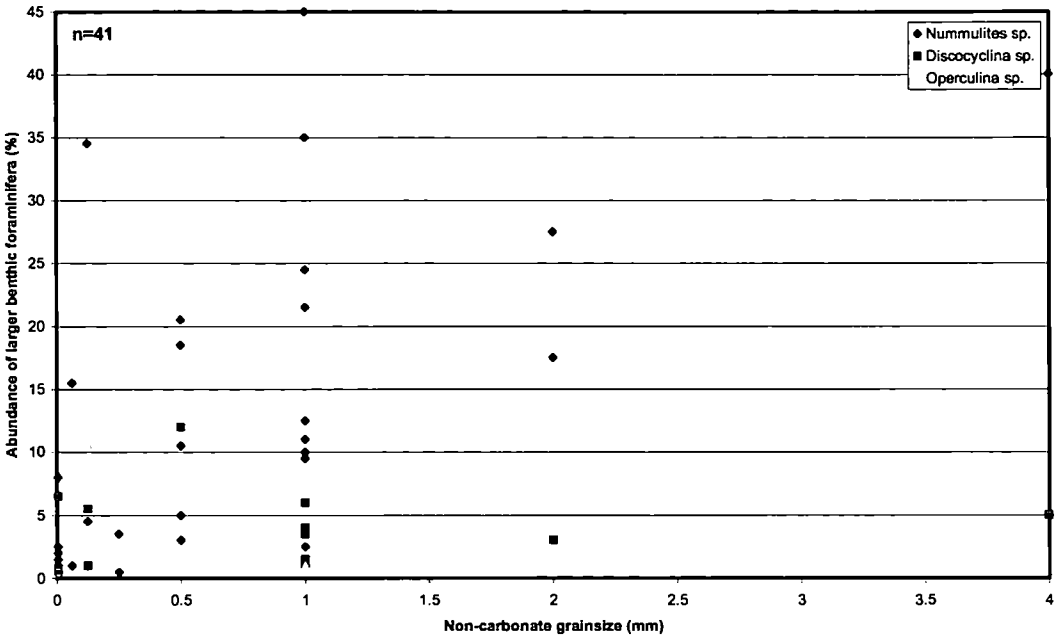


Figure 6.17 Plot of larger benthic foraminifera abundance (%) against median grain size of the largest non-carbonate clasts from selected samples in the Calders area (41 samples).

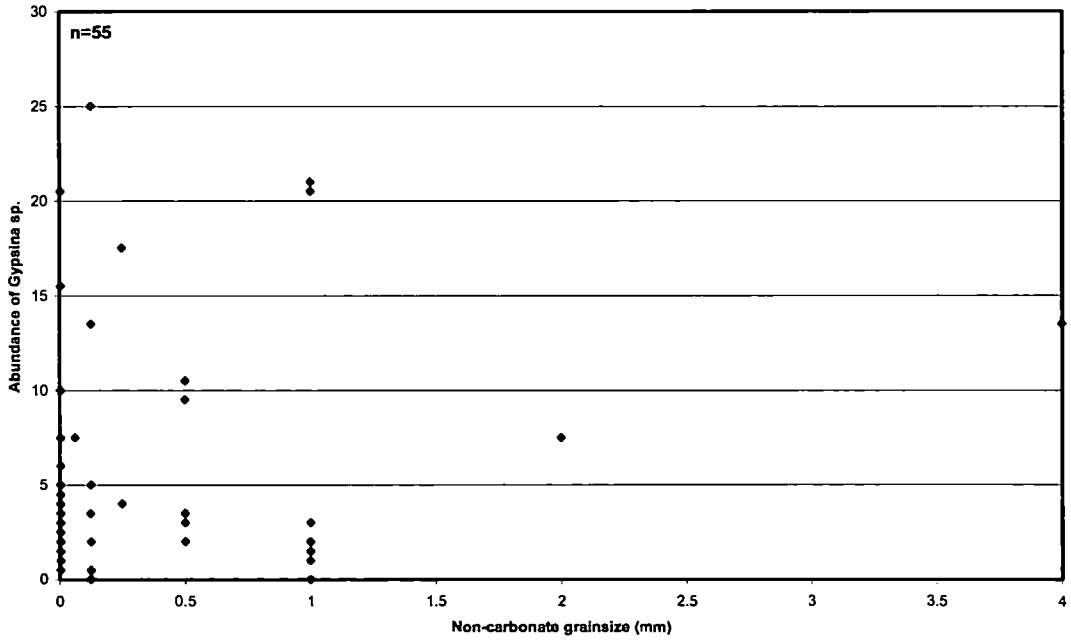


Figure 6.18 Plot of *Gypsina* abundance (%) against median grain size of the largest non-carbonate clasts from selected samples in the Calders area (86 samples).

Corals are particularly abundant where non-carbonate grainsize ranges from 4 to 125 μm (**Figure 6.15**). Fine-grained siliciclastic sediment suspended in the water column will result in increased turbidity with a reduction in the depth of the photic zone and the habitable area of the shelf (**Section 6.1.2**). This will have an effect on the abundance, distribution and morphology of corals. Corals may adapt flattened, platy morphologies as a consequence of low incident radiation (Titlyanov and Latypov 1991), although this adaptation is not advantageous if rates if sediment settling are high. Corals develop branching morphologies in settings characterised by high sedimentation rates where competition for space is high (cf. Grasso and Pedley 1988). Steep, upright faces of branching corals make them less susceptible to sediment settling than planar or concave surfaces. However, steep surfaces will experience a lower degree of incident radiation, thus would not be advantageous in turbid low-light settings. In the Calders area, corals present in lithologies containing very abundant (>25 wt. %) fine-grained, non-carbonate sediment have delicate branching morphologies (**Figure 6.13**). It is postulated that corals within muddy, silicicasltic-rich lithologies of the Calders area developed branching morphologies as a response to high inputs of fine siliciclastic sediment. Fine-grained, non-carbonate sediment would have settled from the water column relatively quickly in a low-energy depositional setting (**Section 3.3.1.4**), thus the water column may have been turbid for short periods of time only. Platy coral development is evident in the second carbonate interval within lithologies containing up to 26 wt. % non-carbonate (**Figure 6.13**). It is inferred that the second interval developed under more turbid conditions, with re-suspension of fine sediment and lower net sedimentation rates.

6.4.2 Larger benthic foraminifera

A summary of the dominant larger benthic foraminifera within the identified facies of the Calders succession, and a definition of their environmental restrictions, is provided in **Appendix 1**.

Most larger benthic foraminifera are dependent upon algal endosymbionts for photosynthesis. This requirement restricts them to the photic zone, which, in clear water settings, may be as deep as 130 m (Ghose 1977, Hottinger 1983, Hallock and Glenn 1986, Murray 1991, Hohenegger *et al.* 1999, Geel 2000). Their occurrence and abundance is controlled by a number of external environmental factors such as

temperature, oxygenation, salinity, nutrient levels, incident light, energy regime and the nature and stability of the substrate (Blondeau 1972, Hallock and Glenn 1986, Racey 1990). Foraminifera test morphology can be related to water depth (and degree of light intensity) and water energy. Thick, robust tests are common within a high-energy regime and/or where the light intensity is high. Conversely, thin tests, often with an increased diameter, occur where the light intensity is lower and/or within a low-energy regime (Hottinger 1983, Hallock and Glenn 1986). Foraminifera with a large diameter also develop in settings where the nutrient supply is sufficient to sustain growth, but does not promote reproduction (Geel 2000).

Larger, free-living foraminifera are typical of muddy substrates (Hallock and Glenn 1986, Racey 1990), and are able to clean their tests using their pseudopodia (Van der Zwann *et al.* 1999). Larger foraminifera mortality may result from burial, as symbiotic algae are prevented from photosynthesising. Some species are mobile and can excavate themselves in the event of shallow burial (Alve 1999).

6.4.2.1 Larger benthic foraminiferal abundance

It is evident from this study that larger benthic foraminifera are able to tolerate prolonged inputs of non-carbonate sediment (**Section 6.3.1**). It is demonstrated that *Nummulites* and *Discocyclina* are able to survive in sediments containing 65 wt. % non-carbonate material, and *Operculina* can tolerate up to 50 wt. % non-carbonate (**Figure 6.10**). Larger benthic foraminifera may be the only biota present in any abundance in siliciclastic sediments with concentrations of 10 to 30 wt. % non-carbonate (**Figure 6.10**). Their abundance may be attributed to a lack of competitors (or predators) in a stressed environmental setting, an ability to survive burial for a short period of time, to excavate themselves following shallow burial and an ability to colonise barren substrates.

Larger benthic foraminifera are not observed where the grain size of the non-carbonate sediment is > 4 mm, and are most abundant at grain sizes ranging from 4 μm to 2 mm (**Figure 6.17**). *Operculina* is not present where non-carbonate grain size > 125 μm. Their absence within very coarse-grained siliciclastic lithologies may be related to the increased abrasion of large grains, particularly in shallow, high-energy settings. *Nummulites* observed along the upper surfaces of conglomeratic units

(**Figure 3.50**) are reworked. The ability of larger benthic foraminifera to modify their test morphology in order to adapt to low light intensities may explain their relative abundance in fine-grained sediments.

6.4.2.2 Morphological response of larger benthic foraminifera to sediment input

Larger benthic foraminifera within the studied sedimentary succession at Calders show a variety of test morphologies. These morphological variations are related to environmental stresses such as siliciclastic sediment input and low light intensity.

Large test diameters in larger benthic foraminifera are usually associated with growth under the influence of low levels of incident light, within a deep-water, low-energy setting (Hottinger 1983, Hallock and Glenn 1985, 1986). The growth of a large test will increase the surface area available for photosynthesis by algal symbionts. A similar morphological response can be triggered in shallow water where an input of clay to silt-grade sediment that remains in suspension will reduce the photic zone. Larger test diameters are associated with low incident light and low energy conditions, thus tests are thin and delicate. Conversely, in high-energy, clear shallow water conditions, foraminifera tests are small with thick walls that limit the amount of ultra violet radiation entering the test, protecting its protoplasm from radiation damage (Hottinger 1983, Hohenegger *et al.* 1999).

The larger foraminifera *Operculina* is found in abundance only in muddy, mixed carbonate-siliciclastic sediments (**Section 3.3.2.3**). *Operculina* tests are large (up to 2.5 cm) and thin with diameter to width ratios up to 0.08. The large diameter of *Operculina* in the studied sediments is related to a reduced photic zone as a consequence of turbidity (not depth as *Operculina* is found within shallow-water sediments-see **Section 3.4.1**). *Nummulites* and *Discocyclina* tests within siliciclastic packstone lithologies are also large with diameters ranging from 1.0 mm to 3.5 cm. The largest foraminifera tests, typically encountered just below the transition from mixed carbonate-siliciclastic to carbonate-dominated sedimentation, demonstrate thick test walls, with diameter to width ratios up to 0.2. Therefore, it is proposed that these large foraminifera test diameters were associated with non-carbonate sediment input but in moderately shallow water with high levels of incident light. Where

environmental conditions are stressful (but tolerable for growth), reproduction of larger benthic foraminifera is delayed until conditions become more suitable (**Figure 6.19**). The stressful conditions in this environment are thought to be related to a combination of sediment input and an energetic hydrodynamic regime.

Rotaline foraminifera present within the carbonate-dominated lithologies, (predominantly *Amphistegina* with relatively rare *Nummulites*) have small, robust test morphologies in response to high incident light. Carbonate-dominated lithologies contain no siliciclastic grains > 1 mm, but may contain up to 38 wt. % clay to silt grade non-carbonate. This grain size fraction could remain in suspension for some time prior to deposition, although no reduction in incident light can be inferred from the morphology of *Nummulites* and *Amphistegina*. It can be concluded from this that although the carbonate lithologies were influenced by a fine-grained, non-carbonate sediment input, but sediment did not remain in suspension for a long enough period to significantly influence the development of larger benthic foraminifera.

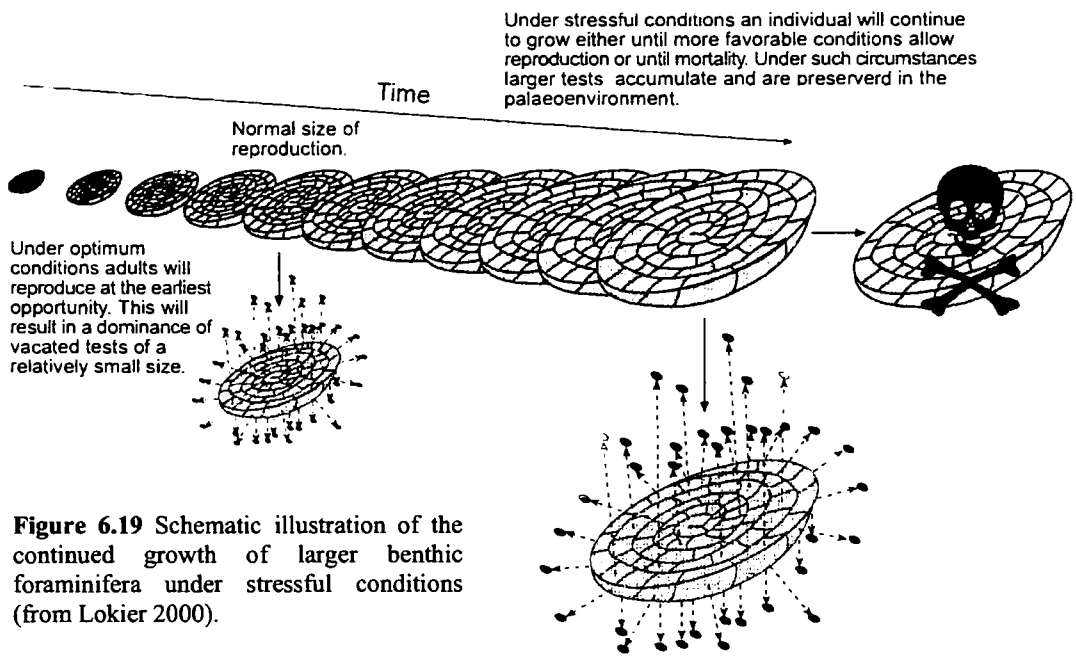


Figure 6.19 Schematic illustration of the continued growth of larger benthic foraminifera under stressful conditions (from Lokier 2000).

6.4.2.3 Sediment re-colonisation by larger benthic foraminifera

The Calders sedimentary succession developed due to repetitive re-colonisation of a siliciclastic substrate by carbonate-producing organisms following a rapid sedimentation event that destroyed the previous benthic habitat and effectively created a new one (**Chapter 3**). *Nummulites,* often followed by *Discocyclina,* are typically the first organism to re-colonise the 'new' habitat in appreciable numbers following an extended input of siliciclastic material. Re-colonisation by larger benthic foraminifera following 100 % mortality of biota is potentially a difficult process, as foraminifera have to reach the new site. Alve (1999) notes that most motile benthic organisms have four different ways of dispersion:

1. The release of gametes, zygotes or embryonic agamonts following sexual and asexual reproduction into the water column;
2. Self-locomotion along the sea floor;
3. Passive entrainment into the water column and subsequent transport of different growth stages;
4. Adaptation to a meroplanktonic juvenile life stage with subsequent passive spread by currents.

Dispersal through the release and transport of embryonic juveniles is thought to be the main mechanism for larger benthic foraminifera (Alve 1999). Benthic foraminifera are not easily entrained while alive as they cling to the sediment with their pseudopodia (Severin and Lipps 1989), and self-locomotion is a very slow process, efficient over very small distances only (centimetres to metres) (Kitazato 1988a). The adaptation to a temporary planktonic life stage is only identified for a few species of smaller benthic foraminifera, and is not thought to be an important mechanism for larger benthic foraminifera (Alve 1999).

The rate of colonisation of a new substrate is related to the hydraulic regime in, and the transit time from, the source area inhabited by species capable of colonizing the new habitat, assuming food and other environmental requirements are not limiting factors (Alve 1999). If high-energy conditions dominate in the source area (bottom current velocities > 20 cm/s) the transit time for more species into the new habitat will

be short (hours to days), colonisation may occur within days and stabilisation may occur within days to weeks. There are no obvious breaks in sedimentation in the Calders area, such as erosion surfaces, hardgrounds or palaeosol horizons, with siliciclastic to carbonate contacts being transitional, thus there was no sedimentary hiatus before colonisation of the siliciclastic substrate that would leave a mark in the geological record.

The first organisms to colonize the new environment will be r-strategists (opportunists) that have high reproduction capacity and short life cycles (Alve 1999). The r-strategists will be succeeded by more habitat-selective, k-strategists. The abundance of *Nummulites,* and to a lesser extent *Discocyclina*, within siliciclastic packstone lithologies that succeed the largely non-fossiliferous sandstones (**Figures 6.4** to **6.6**) is consistent with an initial episode of re-colonisation by r-strategists. These foraminifera are absent throughout the majority of the siliciclastic lithologies, and therefore may have been transported into the area during the embryonic juvenile stage of reproduction (see above). It is postulated that transport of foraminifera in their juvenile stages may have been difficult in this environment as low to moderate-energy conditions are inferred for the Vic Basin from the abundance of carbonate mud (**Chapter 3**). However, the northern and southern margins of the Vic Basin were populated by a large number of indepenent reef systems during the late Middle Eocene (**Chapter 2**), thus potential colonisers had only a small distance (tens of kilometres) to be transported to the Calders area.

6.4.3 Larger encrusting and epiphytic foraminifera

There is a great generic diversity of encrusting and epiphytic foraminifera in the Calders sediments, and there is evidence to suggest that non-carbonate sediment input influenced their abundance through creating and modifying/destroying habitats. Larger encrusting and epiphytic foraminifera identified are *Gypsina, Haddonia, Chapmanina, Fabiania* and victoriellids (*Korobkovella* and *Gyroidinella*). Although major contributors to Recent tropical bioconstructions, these large encrusting foraminifera do not posess photosymbionts and thus their abundance and distribution would not have been influenced by turbidity and light penetration but the nature of the

substrate ie. hard vs. soft (cf. Romero *et al.* 2002). It is postulated that the nature of the substrate would have been strongly affected by non-carbonate input.

The acervulinid foraminifera *Gypsina* is able to tolerate variable amounts of siliciclastic sediment input, and is found to inhabit a variety of carbonate and siliciclastic-dominated habitats in the Calders area (**Chapter 3**). *Gypsina* is found in in sediments containing up to 69 wt. % non-carbonate (**Figures 6.3** and **6.11**), although it is most abundant at non-carbonate concentrations around 2 to 20 wt. %. It is interpreted that in environments characterised by high siliciclastic sediment input and unstable substrates, *Gypsina* adapts to an epiphytic mode of life (**Chapter 3**). This is inferred from the curved nature of *Gypsina* tests. Sea grass meadows, which can develop in such environments, provide adequate stable substrates for epiphytic organisms where the risk of burial is negated (cf. Carbone *et al.* 1994). *Gypsina* is also seen encrusting *in situ* corals (in association with coralline algae) within environments characterised by low to moderate inputs (<20 wt. %) of fine-grained non-carbonate sediment (**Figures 6.5** and **6.6**). Perrin (1992) suggests that *Gypsina* is a common component in reefal environments where there is reduced competition for substrate encrustation, often related to low-light levels. This is interpreted as an indicator of periods of enhanced turbidity at sites of reef growth.

The large encrusting foraminifera *Haddonia, Fabiania, Chapmanina* and victoriellids (*Gyroidinella* and *Korobkovella*) inhabited cryptic environments within coral debris (**Chapter 3**). It is postulated that an input of fine-grained non-carbonate would result in mortality of these organisms through burial (as they are sessile and unable to relocate) and destruction of their preferred habitat through the infiltration of fine-grained sediment into cryptic environments. Large encrusting foraminifera are present in sediments containing up to 54.5 wt. % non-carbonate material, although are more abundant at concentrations < 10 wt. % (**Figure 6.12**). Larger encrusting foraminifera are more abundant where the grain size of non-carbonate sediment is < 1 mm (**Figure 6.14**). *Haddonia* and *Fabiania*, in addition to *Gypsina,* are also important constituents of foralgal rhodoliths (**Section 6.4.4**), and it is proposed that in environments where the risk of burial is high, a free-living mode of life may offer some advantages.

6.4.4 Coralline algae

Coralline algae are important sediment binders and framework builders in reef environments. Coralline algae are photosynthetic organisms and their distribution is controlled by incident light (Adey and Adey 1973, Ghose 1977). The algae grow in the photic zone at maximum depths of 130 m (Tsuji 1993), though a typical depth of < 90 m is more common (Bosence 1993). Coralline algae are common inhabitants of reef and forereef settings in both Tertiary and Recent environments (Adey and MacIntyre 1973, Wray 1977, Ghose 1977) and are an important biotic component of the Calders sediments (**Chapter 3**). In the study area, coralline alage are present as laminar crusts on corals, rhodoliths and articulated forms (**Chapter 3**). The potential influences of silicicilastic sediment input on coralline algae abundance and morphology is demonstrated in sediments in the Calders area.

6.4.4.1 Coralline algae abundance

As described in **Section 6.3.1**, there is a distinctive change in coralline algae abundance in response to non-carbonate sediment input. Coralline algae occur in sediments containing up to 38.5 wt. % non-carbonate material, and are most abundant at concentrations of 10 to 30 wt. % (**Figure 6.9**). Rapid deposition of siliciclastics would result in the burial and mortality of coralline algae, essentially as the alga is a sessile organism and unable to excavate itself if buried. This is suggested from the dearth of coralline algae in siliciclastic-dominated sediments (**Figures 6.3** and **6.9**). In addition, a constant supply of siliciclastic sediments would create an unstable substrate that would be unsuitable for colonisation by most benthic fauna.

The absence of coralline algae in sediments where the grain size of non-carbonate grains exceeds 4 mm, and the rarity of algae at grain sizes between 2-4 mm (**Figure 6.16**) is attributed to the abrasion of soft tissues. Coralline algae are very abundant in sediments where the non-carbonate grain size fraction is between 4-500 μm (**Figures 6.14** and **6.16**). This seems unusual as it is thought that clay to silt-grade sediment, which would remain in the water column effectively reducing light penetration, would reduce the depth of the photic zone. Where algae are no longer able to photosynthesise, mortality would result. Turbidity would be enhanced at moderate energy levels, and turbidity could be maintained through the constant re-

suspension of fine-grained sediment. The success of coralline algae in these muddy settings in the Calders area could be attributed to a number of factors. Firstly, low to moderate energy conditions are interpreted for most of the carbonate-dominated depositional environments (**Chapter 3**), and thus turbid periods may have been of a short enough duration not to significantly affect algae growth. Secondly, clay-grade material may flocculate and thus settle from the water column at lower current energies. Flocculated clays would also be less susceptible to re-suspension. Finally, coralline algae are most commonly encountered encrusting corals, thus they would have occupied an elevated position where the effects of turbidity or sedimentation would be reduced.

6.4.4.2 Changes in coralline algae morphology

Coralline algae occur as a variety of growth forms including articulated, laminar crustose, complex branching morphologies and rhodoliths. It is postulated that sedimentation, and turbidity generated as an effect of non-carbonate sediment input, has influenced coralline algae morphology.

Thin (<0.5 mm), but laterally extensive laminar coralline algae crusts are identified on delicate branching corals in sediments where the amount of non-carbonate exceeds 30 wt. %. Steneck (1986) describes how thin, laterally extensive coralline algae crusts are typical of deep-water, low-energy conditions where competition for space and predation is limited. Turbidity, generated through an input of fine-grained siliciclastic sediments, reduced the incident light, resulting in the development of a deep-water growth form in relatively shallow water. More nodular growth forms characterise relatively high-energy settings with a significant siliciclastic input in the study area (**Figure 6.19**). Nodular growth forms of coralline algae are more common in higher-energy settings and where competition for growth space is high (Minery *et al.* 1985, Minnery 1990).

Rhodoliths are an important constituent of a number of facies in the study area, in particular the coralgal foraminifera rudstone (**Section 3.3.1.5**) and coral bioclastic pack/rudstone (**Section 3.3.1.3**) facies that were deposited in moderately high energy conditions. High-energy conditions may be beneficial for coralline algae growth as fine-grained material is removed from the environment, and it will

encourage the turning of rhodoliths. Rhodoliths in the studied sediments often demonstrate highly complex columnar branching. Concentric laminar and densely branching radial forms tend to develop in higher-energy environments, with more loosely laminated and open branching forms in lower-energy environments (Bosence 1991). A change from large, spheroidal concentric laminar rhodoliths to smaller, more complex branching and columnar forms has been observed in sediments containing a high percentage of clastic material (Lokier 2000).

In unstable sedimentary environments there may be a lack of suitable substrate for sessile encrusting organisms. However, hard surfaces of rhodoliths provide ideal attachment sites for organisms such as foraminifera (cf. Lokier 2000). It has been noticed in this study that incorporated into rhodoliths are large encrusting foraminifera, in particular *Haddonia, Fabiania* and *Gypsina* (**Figure 6.20**). Foraminifera that have encrusted rhodoliths may have been subsequently overgrown by coralline algae resulting in an area of local topographic relief at the rhodoliths surface. Such areas have been amplified through successive growth layers of coralline algae or further encrustation by foraminifera.

6.4.5 Burrowing fauna

Bioturbation is difficult to identify in the carbonate-dominated lithologies due to the recrystallised nature of the sediments, and the very well sorted nature of the micrite matrix. Bioturbation is only identified in carbonate lithologies through subtle colour variations of the matrix and localised concentrations and alignment of bioclasts. Bioturbation is better preserved within siliciclastic-dominated lithologies due to clear grain size variations. Trace fossils present are dominated by *Skolithos* and *Thalassanoides*-type forms orientated perpendicular to bedding. Burrows do not cross bedding contacts. It is inferred from the dearth of bioturbation within the majority of the siliciclastic sediments (**Figures 6.10 to 6.13**) that high-sedimentation rates prevailed, inhibiting colonisation by most burrowing organisms. The only fauna identified within siliciclastic-dominated sediments are irregular echinoids (*Echinolampas*) and large gastropods (*Cerithium*).

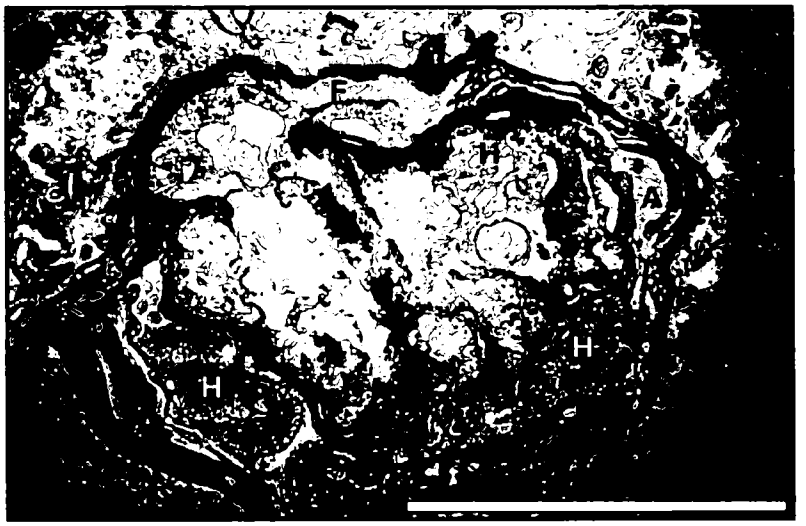


Figure 6.20a (PPL) Example of a complex foralgal rhodolith containing, in addition to coralline algae, *Haddonia* (H), *Fabiania* (F) and acervulinid encruster (A). The nucleus is a coral fragment (sample LCA 99e, log CA-10). Scale bar=5mm.

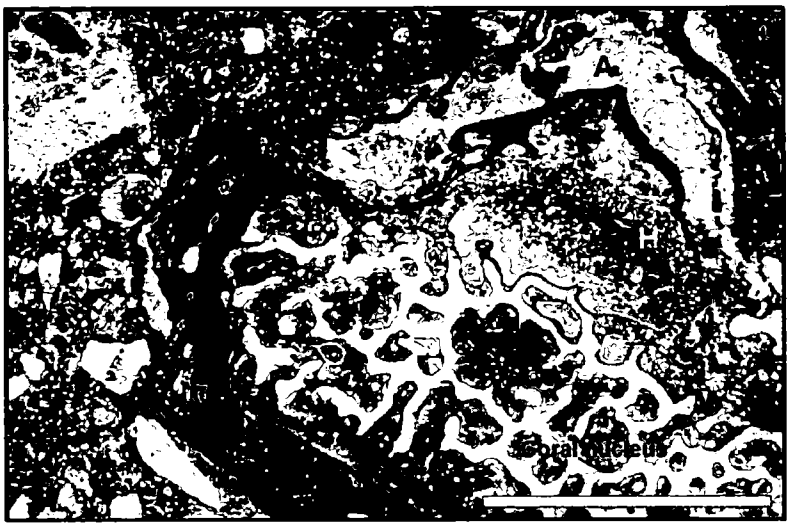


Figure 6.20b (PPL) In some circumstances, foraminifera are observed to encrust the exposed, hard surface of a rhodolith creating an area of localized microtopography. If the rhodolith is not turned frequently, further encrustations will increase the rekief, forming a protuberance (sample LCA 43b, log CA-4a). Scale bar=2mm.

Figure 6.20c (PPL) Warty coralline algae growth within a moderate to high-energy, mixed carbonate-siliciclastic setting (sample LCA 31, log CA-4). Scalebar=2mm.

6.5 Summary

There have been a number of studies investigating the development of carbonates in siliciclastic settings (Santisteban and Taberner 1988, Woolfe and Larcombe 1998, Larcombe *et al.* 2002). However, these studies have not aimed to quantitatively assess the effects of sediment input on the biota present. Studies on modern reefs have concentrated on individual species or organisms, rather the community as a whole (Gleason 1998, Barnes and Lough 1999 and Wesseling *et al.* 1999). Studies quantitatively detailing community change associated with volocanoclastic and siliciclastic input have been presented by Lokier (2000), Wilson and Lokier (2002), Wilson (*submitted*). A summary of the effects of clastic input on biota in the Calders study area is presented below.

6.5.1 Benthic community change

It is expected that a sediment input (or conversely a variation in sediment input) into a shallow marine environment would result in a distinctive change in the benthic community. The punctuated nature of siliciclastic input in the Calders area resulted in the development of two distinctly different environments: prograding clastic shelf and prograding/aggrading coralgal reef with associated detrital sediments. The carbonate accumulations at Calders developed on a periodically abandoned, shallow marine siliciclastic substrate, situated on the southern margin of an east-west trending oceanic inlet. Although the sedimentary regime was siliciclastic-dominated, carbonate development occurred during windows of opportunity provided by abandonment of the siliciclastic substrate or significant reductions in siliciclastic input.

Siliciclastic-dominated beds in the study area are largely devoid of body fossils, indicating unstable substrates with sedimentation rates too high to allow colonisation and habitation. High sediment accumulation rates are confirmed from the presence of vertical and sub-vertical domichnial burrows. Fossils preserved are fragmented and abraded, indicating that they were transported into the environment and/or reworked prior to deposition, and therefore do not represent an *in situ* community. A strategy adapted by some organisms for survival in settings with mobile substrates is the ability to lead an epiphytic lifestyle. Epiphytic forms of

Gypsina are particularly common in siliciclastic rich sediments in the Calders study area. It is inferred that an abundance of encrusting forms of *Gypsina* proliferated in an environment where there was little competition for space.

The switch from high to low/moderate siliciclastic sedimentation rates and early stabilisation of the substrate resulted in the proliferation of a low-diversity benthic community dominated by mobile organisms including larger foraminifera (*Nummulites, Discocyclina* and *Operculina*) and robust, tightly laminar foralgal rhodoliths. Although sedimentation rates were relatively low, the substrate was prone to reworking, as shown by the development of ephemeral channels. This reworking would have inhibited the development of many sessile organisms, including colonial corals, articulated coralline algae and bryozoa.

A further decrease in siliciclastic sedimentation rates, accompanied by stabilisation of the substrate, resulted in the development of a highly-diverse benthic community. This diverse community, dominated by corals, encrusting foraminifera, coralline algae and molluscs, thrived. A constant fine-grained clastic input is identified in the Calders area. However, this background sedimentation was insufficient to bury biota, probably because it remained in suspension through a combination of local and longshore currents, and corals were able to effectively remove sediment from their tissues using one or more of the mechanisms described in **Section 6.1.1**. High/increasing nutrient levels associated with continental runoff led to increased encrustation by coralline algae and a high incidence of bioerosion, particularly by lithophagid bivalves.

6.5.2 Scarcity of fauna within siliciclastic horizons

The switch from carbonate-dominated to siliciclastic-dominated deposition is abrupt. The absence of an intermediate, larger foraminiferal community between the coralgal reefs and shelf clastics in all but one instance indicates that the sediment input was large and rapid. Sedimentation rates were too high to allow colonisation by most sessile benthic organisms, with the rare fauna dominated by burrowing echinoids and soft-bodied burrowing invertebrates.

The relative scarcity of fauna within siliciclastic horizons in the Calders study area is a consequence of the amount and rate of sediment input. Substrate instability

would have inhibited colonisation by corals in their larval stages. Coral larvae cannot remain stable on immobile substrates, and growth will only initiate where the rate of particulate sedimentation is zero or slightly negative (Woolfe and Larcombe 1998, 1999). Larger benthic foraminifera (*Nummulites*) and epiphytic foraminifera are the only biota present in appreciable amounts in siliciclastic-dominated sediments. *Nummulites* was able to survive in this setting through its ability to excavate itself from sediment following burial, and its ability to adapt morphologically to low-light conditions. *Gypsina* was able to survive in a siliciclastic-dominated environment through adapting to an epiphytic mode of life, encrusting sea grass that provided a relatively stable substrate. A summary of the relative abundance of benthic foraminifera encountered in different sedimentary environments is presented on **Table 6.1**.

6.5.3 Preservation potential of carbonate communities

The rapid physical burial of a reef with sediment can preserve an entire community, preventing post-mortem degradation through bioerosion, and physical breakge and transport by currents. The continuity of section exposed at Calders, and the apparent absence of erosion surfaces, hardgrounds etc, suggests that close to the full suite of hard-bodied benthic biota present may have been preserved. It should be noted that some benthic fauna i.e. larger benthic foraminifera, might be able to excavate themselves upon burial, surviving an input of sediment. In such a case, the remaining community will represent a death assemblage or organisms with hard parts, containing only sessile organisms that were unable to excavate themselves rather than representing the true benthic community present.

6.6 Conclusions

In the Calders area of NE Spain, the varying input of clastic material had a strong influence on the development and preservation of the benthic biota. These effects depended upon the organisms present, the amount of sediment input and the sediment grainsize. A divese benthic community, comparable in diversity of modern reefs, existed during times of low siliciclastic input. High rates of sediment input, related to channel avulsion episodes, resulted in the burial and mortality of all sessile

benthic marine biota. This resulted in the establishment of a new environment that was re-colonised by an r-selective oligotrophic to mesotrophic biota dominated by larger benthic foraminifera (*Nummulites, Discocyclina* and *Operculina*) and coralline algae.

7. Carbonate development in siliciclastic environments

7.1 Introduction

Traditionally, carbonates and siliciclastics have been studied separately, likely a consequence of the high degree of specialisation in sedimentology and the different methods of analysis employed to study the two types of deposit. A great deal of information about depositional environments in mixed shelf settings may be acquired from the study of these types of sedimentary successions as a whole. Ancient coral reef systems are useful palaeoindicators of shelf morphology, tectonic activity, local circulation systems, climatic conditions and relative sea-level fluctuations (Leinfelder 1997). The presence of reefs within and adjacent to siliciclastic sediments at Calders (**Chapter 3**) and Altorreal (**Chapter 5**) provides evidence that reefs were also influenced by sedimentary processes of fan-delta systems and siliciclastic shelves.

There is a common misconception that significant shallow water carbonate development is inhibited in areas of active siliciclastic input. However, a diversity of mixed carbonate siliciclastic sedimentary deposits are known from the rock record (Quaternary: e.g. Purser 1987, Iryu *et al.* 1998, Katsutoshi and Seiko 1998; Tertiary: e.g. Santisteban and Taberner 1988, Reinhold 1995, Hayward *et al.* 1996, Wilson and Lokier 2002; Mesozoic: e.g. Nose 1995, Nose and Leinfelder 1997, Leinfelder 1997, Bernecker *et al.* 1999; Paleozoic: e.g. Nield 1982, Malsheimer *et al.* 1996, Long 1997). Additionally, recent work has shown that carbonates accumulate in a diversity of modern siliciclastic settings (e.g. Roy and Smith 1971, Johnson and Risk 1987, Acevedo *et al.* 1989, Tudhope and Scoffin 1994, El-Sammak *et al.* 1997, Leinfelder 1997, Larcombe and Woolfe 1999, Woolfe *et al.* 2000).

The information gained on the development of the mixed carbonate-siliciclastic successions at Calders and Altorreal is here compared with previously published examples in order to review and critically evaluate the temporal and spatial coexistence patterns between carbonate reef systems and siliciclastics. This chapter will review and discuss characteristic features of mixed carbonate-siliciclastic successions deposited under different climatic (**Section 7.2.1**), tectonic (**Section 7.2.2**) and hydrodynamic regimes (**Section 7.2.4**). Autogenic influences will be discussed in **Section 7.2.3**. In addition, this chapter will discuss the palaeogeographic (**Section 7.2.4**) and biotic (**Section 7.3**) attributes of siliciclastic-influenced reefs, and review sequence development (**Section 7.4**). A further aim of this chapter is to

provide a unified model of reef growth and development in different siliciclastic settings, and to comment on the validity of modern reef classification models.

7.2 Controls on the deposition of shallow marine mixed carbonate-siliciclastic successions

Major controls on the deposition of biogenic carbonate successions are climate, tectonics and (relative) sea level changes (Tucker and Wright 1990) (**Figure 7.1**). On mixed carbonate-siliciclastic shelves, these allogenic controls will ultimately determine:

- The grainsize, rate and amount of terrigenous siliciclastic input to the shelf
- Where in the marine system sediment will accumulate (with regards to the availability of accommodation space and relative depth of the photic zone)
- The nature of the accumulation i.e. types of sedimentary facies, facies distributions and sediment stacking patterns

On a local scale, autogenic sedimentary processes intrinsic to the depositional setting such as hydrodynamic regime, biological activity and water chemistry influence facies patterns. In addition to the factors listed above, these autogenic processes may influence:

- Relative depth of the photic zone through presence/absence of turbidity
- Type and diversity of biota
- Nutrient flux
- Salinity of marine waters

This section will review and critically evaluate the allogenic and autogenic controls on the deposition of shoreline-attached mixed carbonate siliciclastic successions using the studied examples from NE and SE Spain (documented in **Chapters 2 to 5**) and relevant examples. It should be noted that all of the processes above interact, producing a complex range of effects on carbonate development. A specific effect may be the result of a complex interaction of different processes, thus it is not always possible to establish a single cause for an effect when studying

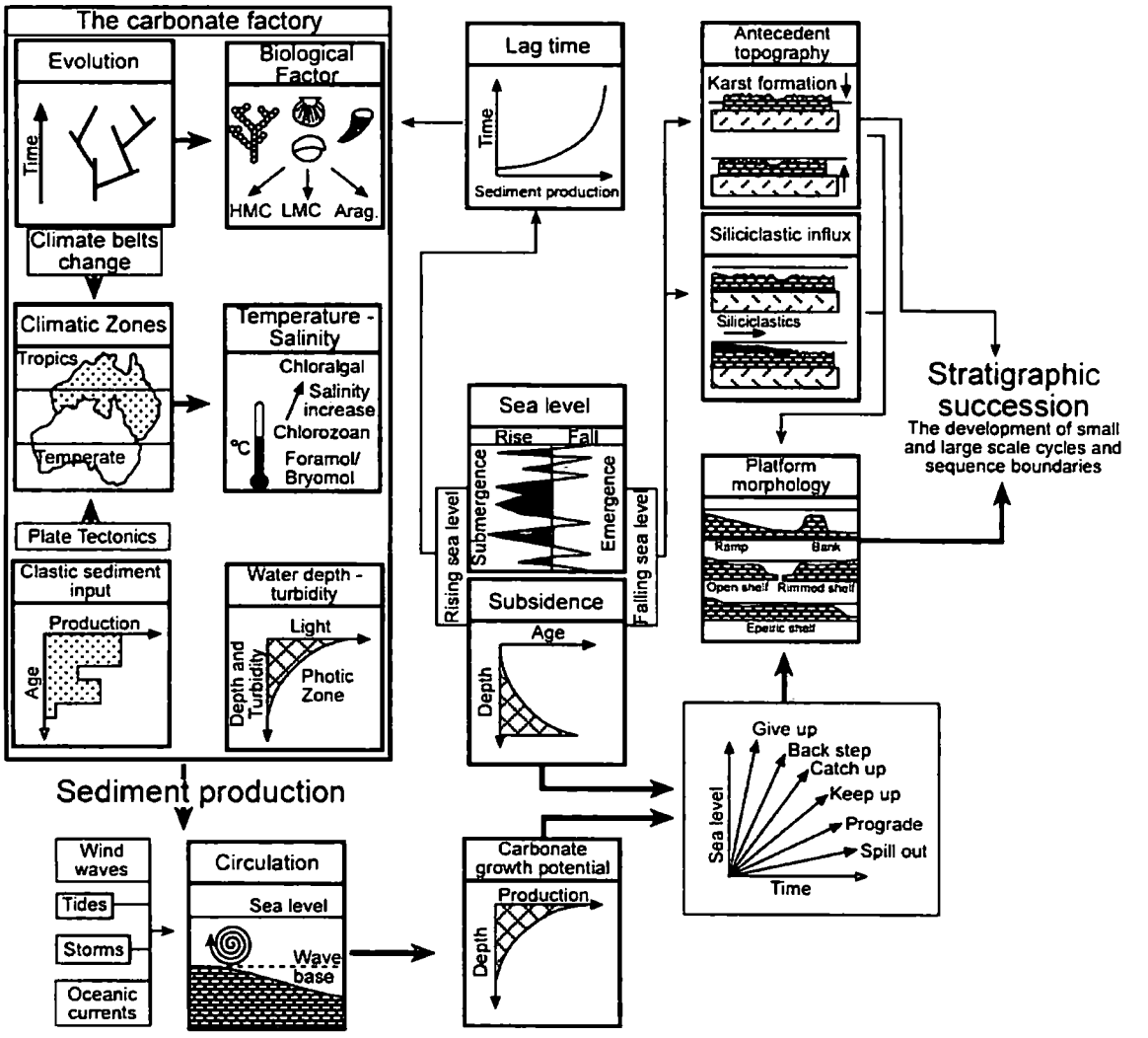


Figure 7.1 Diagram summarizing the main factors controlling the formation of carbonate successions (from Jones and Desrochers 1992)

carbonate depositional systems. In mixed systems establishing a single 'cause' for an 'effect' is even more difficult since the two systems respond differently to allogenic and autogenic influences.

7.2.1 Climate

"The two main aspects of climate are temperature and precipitation. Absolute temperature and precipitation rates are important, but of equal importance are their fluctuations on a seasonal and non-seasonal scale, and the magnitude and frequency of extreme events" (Reading and Levell 1996). Climate is a major control on the frequency, magnitude and grainsize of terrigenous flux to continental shelves. Climate influences the grainsize of siliciclastics transported to marine environments through the degree of physical and chemical weathering of grains, producing fine silt/clay-grade material. This in turn exerts control on the nature of siliciclastic substrates in areas proximal to siliciclastic input i.e. soft and muddy vs. coarse and relatively hard. In addition, the climatic regime may influence marine water chemistry. High precipitation rates and continental runoff can influence nutrient levels, pH and salinity (see **Section 6.1.3**).

The influences of climate (as sediment grainsize, the frequency and magnitude of sediment input, and precipitation and continental runoff rates) on biogenic shallow marine carbonate development are summarised in the following. The influence on biota has been discussed in **Chapter 6**.

1) Grainsize: Where the siliciclastic input is very fine-grained, it may remain in suspension in the water column for a considerable period of time, prolonging turbidity, reducing light penetration and reducing the depth of the photic zone (**Section 6.1.2**). This process will reduce the habitable depth for photoautotrophs. In comparison, where the siliciclastic input is coarse-grained, sediments rapidly settle out from the water column, and the water may clear within a relatively short period of time (**Section 6.1.2**). Excluding complete burial of biota, the most significant effect of coarse input would be abrasion of soft tissues and skeletal breakage (see **Chapter 6**). The grainsize of a potential substrate influences reef initiation. In the Red Sea, gravelly sediments provide an ideal substrate for coral planulae to settle on and

develop. Corals grow directly on non-lithified pebble clasts greater than 60 mm in diameter (Hayward *et al.* 1996). Material finer than this is not normal colonised.

2) Frequency of sediment input: In areas of continuous siliciclastic sedimentation, settlement of photoautotrophs such as corals in their larval stage may be inhibited by an unstable substrate (see **Chapter 6**). It follows that in more humid, equatorial regions characterised by a constant influx of fine-grained, terrigenous sediment, coral recruitment may be difficult in the absence of a pre-existing, stable substrate such as exposed bedrock, pre-existing skeletal build-up etc. In contrast, windows of opportunity for settlement and carbonate production occur during periods of minimal siliciclastic input in semi-arid environments characterised by punctuated siliciclastic input (providing other criteria such as correct temperature and water chemistry are fulfilled).

3) Magnitude of sediment input: High rates of sedimentation may result in the burial and ultimately the death of sessile carbonate producing organisms such as corals, coralline algae and bryozoa (see **Chapter 6**). Organisms within soft-bottomed, unstable environments may have adapted a degree of mobility, allowing self-extraction if buried (Heikoop *et al.* 1996). However, even motile benthic organisms are vulnerable to physical burial if the amount of sediment input is large.

4) Precipitation and continental runoff: An input of freshwater into a marine system in conjunction with an input of terrigenous siliciclastic material can result in rapid changes in water chemistry, including changes in pH, salinity and nutrient levels. High nutrient levels are generally unfavourable to carbonate systems (Schlager 1992, Hallock 1988). Corals and larger benthic foraminifera are oligotrophic (Hallock and Schlager 1986, Schlager 1992, Geel 2000), although some genera may adapt to heterotrophy in if nutrient input is high (Antony 2000, Antony and Fabricus 2000). The growth of fleshy algae, which may encrust and effectively suffocate benthic communities, is stimulated by an increase in nutrient availability (Hallock and Schlager 1986, Hallock 1988). Additionally, an input of freshwater in the marine system may cause fluctuations in marine water salinity in nearshore areas. This may have detrimental consequences for stenohaline organisms that cannot tolerate such

fluctuations (Lees 1975). It should be noted that rates of continental run-off are a function of the hinterland topography, size of the drainage area, vegetation cover and the effectiveness of sediment transport, in addition to the climate.

Siliciclastic sediment flux to near-shore and shoreline-attached depositional systems may vary between humid equatorial settings and arid subtropical settings due to the differing precipitation. In humid equatorial regions sediment input is roughly constant (although many equatorial regions have a seasonal climate). In comparison, in sub-tropical/temperate semi-arid regions, sediment flux to the coastal system may be punctuated but of high magnitude with short-lived, catastrophic events (Hayward 1985, Ahmed *et al.* 1993, Wilson and Lokier 2002).

7.2.1.1 Tropical humid to subtropical temperate sediment fluxes

In humid equatorial to subequatorial regions, precipitation and runoff are moderately high and near-shore or shoreline-attached carbonate systems may be affected by near continuous terrigenous siliciclastic input (Wilson and Lokier 2002). Chemical weathering and long sediment transport pathways yield fine-grained sediments. Many studies investigating the transitions between fine siliciclastics and carbonates found modern reef systems to be many kilometres offshore where water clarity is high (Flood and Orme 1988, Friedman 1988, Murray *et al.* 1988).

Siliciclastic input to the Calders area during the late Bartonian was punctuated, yielding sharp based, muddy sandstone, siltstone and shale beds typically 50 cm in thickness (**Figure 3.38c**), often forming clinoform geometries (**Figure 3.38a** and **b**). Siliciclastic environments were characterised by unstable shifting sandy substrates, which effectively inhibited colonisation by sessile benthic organisms (**Section 3.3.4.1**). Sandstones are muddy, and it is inferred that the clay to fine silt fraction remained in suspension for some time, reducing water clarity, suppressing the photic zone thus inhibiting photosynthesis and the proliferation of photoautotrophs. Siliciclastic facies were inhabited mainly by large *Cerithium* gastropods, echinoids and some larger benthic foraminifera (mainly *Gypsina* if sea grasses were present) (**Section 3.3.1.8**). The biota assemblage of the Calders carbonates includes a range of organisms that are typical of warm latitudes (**Table 7.1**). No climatic data concerning the Vic Basin have been published, although late Tertiary climatic conditions of the Igualada Basin, neighbouring the Vic Basin, are well constrained. During the late

Middle Eocene (Bartonian-Priabonian), the dominance of herbaceous open vegetation reflects a seasonal climatic regime (Cavagnetto and Anadon 1996). The inferred implications of this would be a punctuated sediment input, reflecting variations in precipitation and continental runoff into the Vic Basin on an annual/monthly time scale.

Organism	Temperature range (°C)
Amphistegina	Generally > 20°
Operculina	25° +/- 3°
Hermatypic corals	> 18 °

Table 7.1 The temperature ranges of organisms within the Calders depositional succession (data from Blondeau 1972, Murray 1991)

Coral reef systems developed under siltation stress in the Lower Jurassic Saxony Basin, Germany. At this time, the basin was situated at a palaeolatitude of 39°N within the warm-temperate subboreal realm, although the climate was humid, with distinct wet and dry seasons (Bertling 1997 and references within). The inner shelf was affected by seasonally high continental runoff, with sporadic fine-grained siliciclastic influxes resulting in coral mortality (Bertling 1997). Phases of reef proliferation coincided with periods of reduced run-off, although the amount of suspended sediment and the persistence of turbidity was influenced by the local hydrodynamic regime. Permanently turbid/low light conditions are inferred from the dominance of platy and sheet like coral colonies (Bertling 1997).

7.2.1.2 Subtropical arid sediment fluxes

In comparison to warm and wet settings, terrigenous sediment in arid settings is often locally supplied to the coastline via ephemeral wadis or fan systems. In these arid areas, sedimentation events are often short-lived, of high-magnitude and are typically related to storms (Hayward *et al.* 1985, Purser 1987, Rheinhold 1995). Between major sedimentation episodes, prolonged periods of minor siliciclastic input allows colonisation by photoautotrophs, assuming other environmental conditions

such as temperature, salinity and nutrient levels, are suitable. As a result of localised and punctuated input, reefs develop in clear waters, and deposits may contain little, if any, siliciclastic material.

There are no published data at present concerning climate during the Late Miocene of the Fortuna Basin, although there is a wealth of indirect sedimentary evidence from the lithoclastic facies of the Altorreal study area described in **Section 5.2.1.1**, and the coeval Mediterranean basins of the Spanish Betic region (see **Chapter 4**). Semi-arid conditions are inferred from the absence of chemical weathering of grains and the coarse grainsize of sediments (see **Section 5.2.1.1**). Sedimentation was from discrete, high magnitude events, with sediment locally derived from uplifted Internal Zone massifs (**Chapter 5**). Although the periodicity of sedimentation 'events' cannot be established with the available data, incipient soil development, inferred from the haematite-staining of siliciclastic grains, provides evidence for prolonged subariel exposure of lithoclastic beds between sediment pulses (cf. Mather 1993).

Large oysters encrusted the upper surfaces of boulders between phases of siliciclastic sediment input (**Figure 5.7b**). When the time between successive sediment pulses was sufficient to allow coral colonisation, small *Porites* colonies developed (**Figure 5.7a**). Carbonate production at Altorreal (in the form of coral-microbial reefs) was rarely synchronous with siliciclastic input, evidenced from the virtual absence of siliciclastic material within the reef framework (**Section 5.2.3**). The exception to this was in proximal areas, where coarse siliciclastics supported the narrow, branching *Porites* colonies (**Section 5.2.3.3**). Generally, biogenic carbonate developed on (temporarily) inactive portions of the fan delta system, with reef demise partially attributable to physical burial as the fan was reactivated and sedimentation resumed (**Section 5.3**). The presence of microbial structures (**Figures 5.17c** and **d**) and the evidence for reef erosion prior to siliciclastic deposition (**Figures 5.16b** and **c**), indicate that a number of factors in addition to the sedimentation and the climatic regime were responsible for the demise of the Altorreal patch reefs. Stromatolites are perceived as indicators of enhanced salinity (Kendall and Skipwith 1978, Riding *et al.* 1991, Steneck *et al.* 1998) but at Altorreal, stromatolite development was at least partly synchronous with coral growth (**Section 5.3.2**). Hardy corals such as *Porites* have relatively wide tolerances for salinity variations

(Braithwaite 1971, Downing 1985) but it is unlikely enhanced salinities were responsible for microbial growth, mainly due to the modest faunal diversity. Stromatolite development is related to enhanced nutrient input, a consequence of terrestrial runoff during storm events. Evidence is provided by the temporal and spatial occurrence patterns of stromatolite with respect to corals and lithoclastic facies, discussed in **Section 5.4.3**. It is postulated that there were punctuated influxes of fresh, nutrient-rich water as runoff to the marginal marine environment, even during times of delta lobe abandonment.

Coral reef development in the Miocene Mediterranean Basins (**Table 7.2**), and the Miocene to Recent of the Red Sea region, is comparable to that of Altorreal (in terms of both climate and sedimentation style). Reefs have developed in close association with coarse siliciclastics in the Gulfs of Suez and Aqaba since the Miocene. Mean annual rainfall in the Gulf of Suez is <10 mm yr^{-1} (Friedman 1968). Sediments are locally derived, and are transported to the coast via ephemeral fluvial and wadi systems where they are rapidly deposited in near-shore environments (Hayward *et al.* 1985, Purser 1987). Sedimentation occurs as catastrophic, high-magnitude but infrequent, short-lived events of 1-2 days duration (Hayward 1985, Purser *et al.* 1987, Ahmed *et al.* 1993, El-Sammak 1997). As was deduced for the Altorreal study area (**Section 5.3**), areas of fans are inactive at any point in time. These areas are typically sites of carbonate development (Hayward 1985, Ahmed *et al.* 1993, El-Sammak 1997).

Sporadic sediment influx, deposition in marginal marine environments and intermittent reef growth is described from a number of Miocene basins in southeastern Spain where a subtropical, semi-arid climate prevailed. Examples include the San Miguel de Salinas Basin (Rheinhold 1995), the Granada Basin (Braga *et al.* 1990), the Almanzora Corridor (Saint Martin *et al.* 1989) and the Lorca Basin (Wrobel and Michalzik 1999). Sediment is locally sourced from uplifted Internal and External Zone Massifs, and transported to the coastline via alluvial fans, fan delta and braid delta systems. As described in the Fortuna Basin (see **Chapter 5**), and the examples from the Red Sea, reefs developed on temporarily inactive portions of these siliciclastic sedimentary systems.

A potential consequence of high evaporation rates in arid environments, particularly in (partially) land-locked basins such as the Fortuna Basin, is increased

salinity of marine water. Land-locked basins would be particularly susceptible to variations in water chemistry. This may lead to the development of evaporites. Evaporation rates in the Gulf of Suez range from 2400 to 4250 mm yr^{-1} (Assaf and Kessler 1976). Consequently, evaporites developed in marine structural depressions and with sabkha development onshore (Purser 1987). Evaporites are not present in the Altorreal section, although extensive evaporite deposition occurred within eastern portions of the Fortuna Basin in the late Miocene (Garces *et al.* 2001, Rouchy *et al.* 2001). By the late Tortonian, the Fortuna Basin was semi-restricted. It is postulated that significant input of terrestrial runoff into marginal marine system would have an exaggerated effect on seawater chemistry, in particular water salinity and nutrient levels. The most distinctive indicator of 'difficult' environmental conditions in the studied sections at Altorreal is the presence of stromatolitic structures (discussed above).

7.2.1.3 Summary

The sites and duration of reef development at Calders and Altorreal were influenced to variable extents by the climatic regime, which in turn was an influence on the rate, magnitude and composition of sediment input.

In subtropical, semi-arid settings, such as the Altorreal study area, terrestrially derived sediment is supplied locally to the coastline via ephemeral fluvial or wadi systems during infrequent storm events. Sediments have short transport pathways, and are typically coarse-grained. Reef systems develop in near-shore clear waters during quiet periods between storm events with little siliciclastic interruption. In comparison, nearshore reefs within temperate and humid settings where runoff and precipitation rates are high, such as the Calders study area, may be characterised by near-continuous freshwater and siliciclastic input. Sediments may be finer grained because of longer transport pathways and chemical weathering.

7.2.2 Tectonic setting and sediment supply

"Tectonism affects sedimentation in a number of different ways and on many different scales...movement along faults, growth of folds, block tilting, differential subsidence and uplift provide a critical and delicate control on the type, thickness and distribution of sedimentary facies" (Reading and Levell 1996). In addition to influencing the siliciclastic input to the marine realm, the tectonic regime also

influences shelf morphology that in turn will influence the sites and geometry of shallow carbonate deposits. This is discussed in **Section 7.2.4.2**. Tectonism may lead to the development of physical barriers that can modify water circulation patterns. Changes in circulation patterns may modify seawater chemistry (i.e. nutrient supply, salinity) and the velocity and direction of local water flow. The importance of water chemistry on reef organisms is reviewed in **Section 6.1.3**. Since it is the aim of thesis to evaluate the effects of sediment input on carbonate development, the effects of tectonism on water circulation patterns are only discussed briefly in this section.

Modern and ancient mixed carbonate-siliciclastic successions accumulated under a range of tectonic regimes (**Table 7.2**). Tectonic settings with common siliciclastic input include 1) extensional basins, some with a degree of transtension or transpression, 2) passive margins and 3) foldbelts. Volcanoclastic input associated with carbonate development is common in many tectonic settings; this is not covered here since it has been discussed in detail by Lokier (2000) and Wilson and Lokier (2002). The relationship between tectonic environment and effect on reef development is discussed in the following sections.

7.2.2.1 Extensional settings

Tectonic activity is an important allogenic control on sedimentation, and is most important for timescales $>10^6$ years. Smaller-scale tectonism, such as movement along faults has a strong influence on sediment input in rift systems or compressive plate margins, some with a strong strike-slip component. The development of carbonate provinces in close proximity to siliciclastic sediments in strike-slip settings and graben zones is common within the Miocene of the western Mediterranean (including the Fortuna Basin-this study). Extensive Miocene to Recent reef systems of the Red Sea (including the Gulfs of Aqaba and Suez) also developed in extensional settings characterised by block faulting (Purser 1987, Purser *et al.* 1996).

Table 7.2 Carbonate development in association with siliciclastics within different tectonic regimes

Tectonic setting	Basin/area	Age	Relationships between carbonates and siliciclastics	References
EXTENSIONAL	Western Mediterranean: Fortuna Basin – **THIS STUDY** Lorca Basin San Miguel ed Salinas Basin Guadix Basin Granada Basin	Miocene	Carbonate development on abandoned fan delta lobes	Santisteban and Taberner (1988) Lonergan and Schreiber (1993) Wrobel and Michelzik (1999) Rheinhold (1999) Soria *et al.* (1999) Martin *et al.* (1989, 1992)
	Red Sea, Middle East (including Gulfs of Aqaba and Suez)	Miocene – Recent	Carbonate development on abandoned alluvial and fan delta lobes. Localised reef development on tectonic highs (horsts) with trapping of siliciclastics in structural depressions.	Hayward (1985) Purser *et al.* (1987, 1996) Ahmed *et al.* (1993) El-sammak *et al.* (1997)
	Malta Graban (Eastern Mediterranean)	Miocene	Reef development on the shelf break of slope on a fault scarp. Interruption of reef growth during tectonically induced siliciclastic sedimentation phases.	Pedley (1996)
	Lusitanian Basin, Portugal	Jurassic	Periods of reef development on syn-tectonic marginal marine fan deltas. Reef systems destroyed during phases of tectonism.	Leinfelder (1987, 1994, 1997)

Table 7.2 (cont.)

Tectonic setting and typical characteristics	Basin/area	Age	Relationships between carbonates and siliciclastics	References
FORELAND BASIN AND ASSOCIATED PASSIVE MARGIN	SE Pyrenean Foreland Vic Basin – **THIS STUDY**	Middle Eocene	Carbonate development on abandoned delta lobes. Carbonates affected by re-suspended silciclastics from adjacent, non-carbonate, areas.	Taberner (1983) Santisteban and Taberner (1988) Taberner *et al.* (1999)
	Alpine Foreland Basin, France/Switzerland	Middle Eocene	Carbonate development on the forebulge during a transgression. Carbonates largely unaffected by siliciclastic input, with siliciclastics trapped in the fore-deep.	Sayer 1995
	Kasaba Basin, SW Turkey	Miocene	Carbonate development on inactive portions of fan deltas proximal to the orogenic wedge.	Hayward *et al.* 1996
	Saxony Basin, Germany	Lower Jurassic	Carbonates influenced by sporadic, fine-grained siliciclastic inputs related to enhanced continental run-off during the wet-season	Bertling (1997)
FOLD BELTS	Central Mediterranean and Orania, Algeria	Miocene	Reef development on submerged, thrust-generated bathymetric highs. Sites offered some protection from silcicilastic input.	Pedley (1996) Saint-Martin (1996)

The style of marginal marine siliciclastic sedimentation in the Miocene of the Mediterranean region is often dominated by punctuated inputs of poorly sorted lithoclastic sediments (Martin *et al.* 1989, Braga *et al.* 1990, Rheinhold 1995, Saint-Martin and Cornée 1996, Purser *et al.* 1996). Individual episodes of siliciclastic input are often related to flash floods (see **Section 7.2.1.2**), although these sediment pulses can also be related to tectonic pulses (Lonergan and Schrieber 1993). Tectonism may adversely affect reef development through stimulating debris flows, which may lead to reef burial. Substrate instability due to protracted tectonism also inhibits reef development. An indicator of tectonism would be the presence of metre-scale allochthonous blocks within a sedimentary succession (Lonergan and Schreiber 1993), although these features may also occur in extreme flash floods or from cliff collapse (García-Mondéjar and Fernández-Mendiola 1993). In addition, tectonic uplift of basement provides a ready source of material that may be eroded and reef burial occurs at a later date.

The Fortuna Basin developed within a large left-lateral shear zone extending from Alicante in the northwest to Almeria in the southeast (Sanz de Galdeano and Vera 1992, Montenat and Ott d'Estevou 1990, Montenat *et al.* 1987) (**Chapter 4**). The northern and southern margins were faulted and tectonically active throughout the Neogene (Montenat *et al.* 1987, Montenat and Ott d'Estevou 1990, Sanz de Galdeano and Vera 1992). The present southern margin of the Fortuna Basin is bounded by the Alhama de Murcia Fault, located just to the north of Murcia and relatively close to the site of reef development at Altorreal (**Figure 4.6**). Early in the development of the basin-fill, the overall tectonic regime was extensional. However, from the Tortonian onward, north-south to northwest-southeast compression of southeastern Betic region led to inversion of extensional structures associated with the Alhama de Murcia fault, and the creation of topographic highs (Garcés *et al.* 2001). These topographic highs, composed of Betic Internal Zone lithologies, would have provided an adequate sediment source feeding fan delta systems on the basin margins (Santisteban and Taberner 1988, Garcés *et al.* 2001).

There is no direct evidence for fault-controlled sedimentation at Altorreal, such as growth fault structures or soft sediment deformation. However, short sediment transport pathways and rapid deposition of clasts from cohesive mass flows is indicated from the textural and compositional immaturity of beds (cf. Nemec and

Steel 1984, Mather 1993) (**Section 5.2.1.1**). Sediments are sourced from an uplifting Internal Zone basement block and were transported to the study area at Altorreal through infrequent but high magnitude flash flood events. Internal Zone massifs were largely eroded in the Neogene but are preserved locally as the Orihuela and Carroscoy Massifs to the northeast and south of Altorreal respectively (**Figure 4.6**).

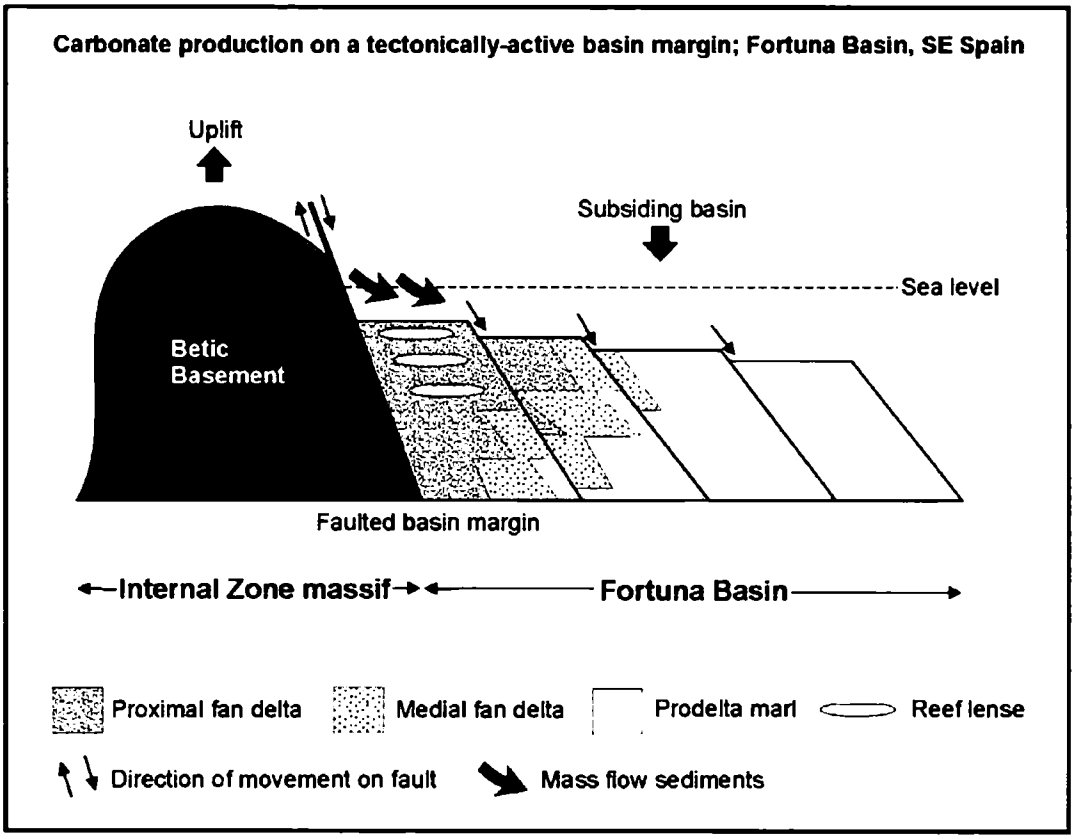


Figure 7.2 Stylised illustration of fault-controlled sedimentation on the western margin of the Fortuna Basin during the Tortonian. Patch reef development occurred on the hanging wall during tectonically-quiescent periods. After Lonergan and Schreiber (1993)

Fault-controlled sedimentation is observed on the western margin of the Fortuna Basin. Mesozoic Betic basement rocks within the Sierra Espuña are juxtaposed against Late Miocene fan delta sediments that are intercalated with coralgal and *Porites* reefs (Lonergan and Schreiber 1993) (**Figure 7.2**). Syn-sedimentary extensional faulting along the basin margin occurred as a consequence of rapid uplift of basement rocks and coeval subsidence of the Fortuna Basin in the late Tortonian (Lonergan and Schreiber 1993). Clasts of fan-delta deposits, comprising Permo-Triassic red beds, dolomites and Mesozoic platform carbonates, are sourced

from the footwall. Movement along the fault triggered mass flow events, depositing steeply inclined conglomerate beds with little internal structure and large allochthonous blocks (Lonergan and Schreiber 1993). Reefs developed on the extensional fault blocks during times of tectonic quiescence, and were influenced little by siliciclastic input. Reef demise in this locality is related largely to resumed tectonism (i.e. movement along the basin-bounding fault) and rapid burial by mass flow deposits (Lonergan and Schreiber 1993). The sedimentary succession is comparable with that observed in the southern Fortuna Basin in the Altorreal area (**Chapter 5**), with inclined, poorly sorted lithoclastic beds often containing shelly debris. However, the presence of boulders in excess of 2.5 metres in the western basin margin suggests that there was either a higher magnitude tectonic influence at the Sierra Espuña locality and/or that the Altorreal succession was deposited further from an active fault.

Evidence for syn-tectonic reef development in the western Mediterranean is evident in the Miocene of Malta. The Ghar Lapsi reef complex developed in association with the tectonically active Malta Graben (Pedley 1996). Block faulting in the region produced a stepped shelf margin with coral reef development largely restricted to the shelf break of slope on a synsedimentary fault scarp (Pedley 1996). Periodic physical disruption of reef development due to earthquakes is inferred from the preservation of metre-scale allochthonous blocks, comparable to those described in the western Fortuna Basin by Lonergan and Schreiber (1993).

Assuming depositional conditions are favourable, phases of coral reef development are likely to occur during periods of tectonic quiescence. Coral reef development on the Kimmeridgian Castanheira fan delta in the Lusitanian Basin, Portugal, are correlated with phases of tectonic quiescence, yielding sedimentary successions very similar to those of the western Fortuna Basin (Leinfelder 1994, 1997). Reef systems developed in association with syn-tectonic marginal marine fan deltas. Reefs were destroyed when there was a resurgence of tectonic activity, buried by significant thicknesses of amalgamated debris flow and collapse breccia deposits (Leinfelder 1987, 1994).

In summary, the generation of topographic highs in settings dominated by block faulting affects reef development in two ways. Firstly, faulting may lead to reef destruction through seismic activity, triggering debris flows and leading to reef burial.

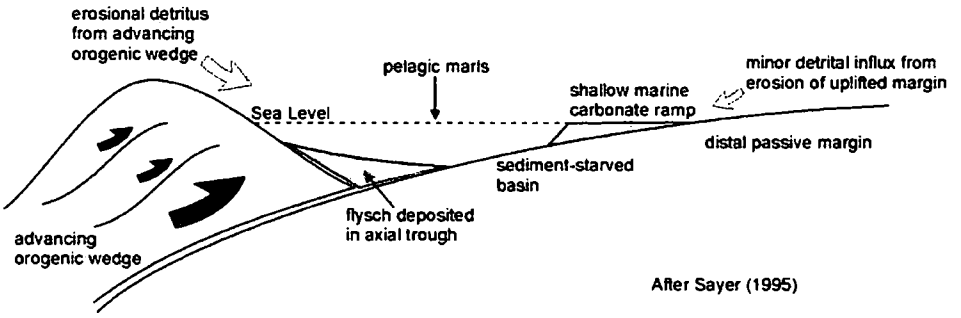
Secondly, the generation of local topographic highs provides sources of sediment that over time are eroded and transported to the coastline.

7.2.2.2 Foreland basins and associated passive margins

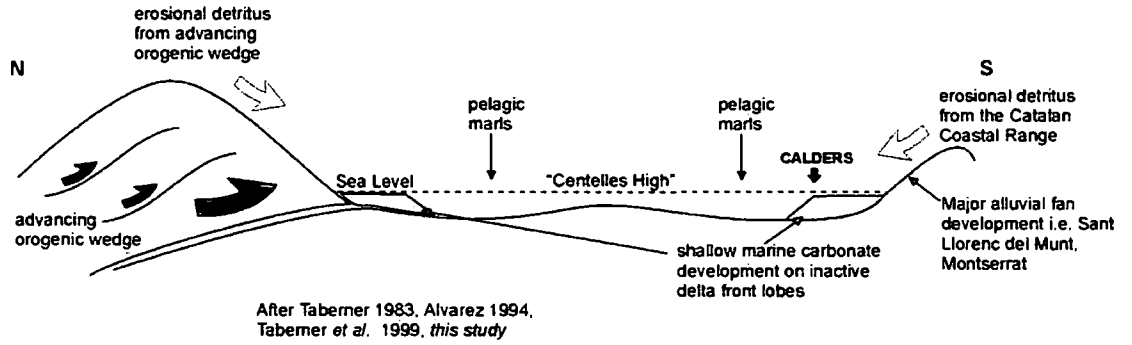
The development of extensive carbonate provinces in foreland basin settings is common throughout geological time. Study of foreland carbonates has largely been focussed on syn-orogenic carbonate platforms that developed during the earliest stages of foreland subsidence and transgression such as the Nummulitique carbonates of the Alpine Foreland Basin (Sayer 1995). Carbonate ramps developed on the fore-bulge during the underfilled stage of basin development when siliciclastic material shed from the advancing orogenic wedge was trapped in the proximal foredeep. Siliciclastic detritus was prevented from reaching the carbonate system and reducing productivity (**Figure 7.3a**). Carbonate development in these settings is influenced by the flexure of the plate (and thus the extent of the cratonic margin within the photic zone), migration of the fore-bulge through continued collision, relative sea-level changes and siliciclastic input (Dorobeck 1995, Sayer 1995).

Carbonate development on the distal margin of a foreland system in the late underfilled stage of basin development is considered unusual due to high siliciclastic input derived from erosion of the passive margin and the encroaching orogenic wedge. In the late Middle Eocene (Bartonian), Calders was located on the southern passive margin of the Southeastern Pyrenean Foreland Basin (**Chapter 2**). The Calders succession is part of the Marine Sequence, the final open marine sediments deposited in this part of the foreland prior to continentalisation of the basin (**Section 2.5.2**). Significant siliciclastic input to the southern margin of the Pyrenean foreland in the vicinity of the Vic Basin has been attributed to the reactivation of Mesozoic fault systems in the Catalan Coastal Range during the Alpine Orogeny (Lopez-Blanco 1993, Vergés *et al.* 1998) (**Section 2.3.4**). The Catalan Coastal Range was a topographically positive feature in the Tertiary, shedding siliciclastic sediment into the basin via major alluvial fan systems such as Montserrat and Sant Llorenç del Munt (Monstad 2000). The Centelles Formation, within which the carbonates of the studied sections at Calders are preserved, is the basinward equivalent of these coastal fan systems (**Figure 7.3b**).

a) Sites of syn-orogenic carbonate development in the underfilled stages of a foreland basin system



b) Sites of carbonate development in association with siliciclastic sedimentation in the late underfilled stage of Pyrenean foreland development



c) Sites of syn-orogenic carbonate development in a proximal foreland setting, Kasaba Basin, SW Turkey

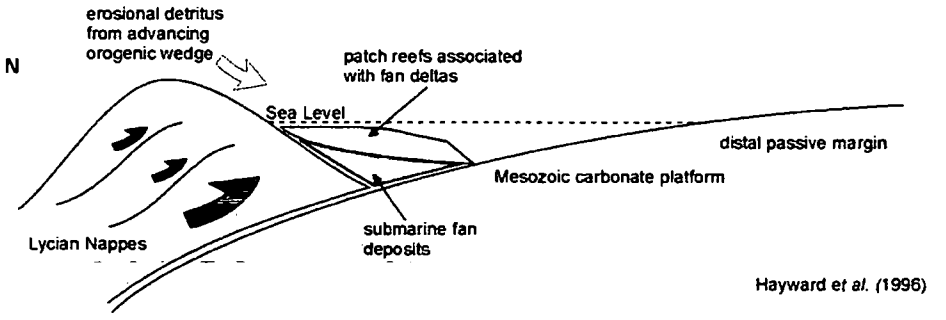


Figure 7.3 Carbonate development in association with siliciclastic sediment input in foreland basin settings. Carbonates typically develop during the under-filled stage (a) where most of the erosional detritus from the advancing orogenic wedge is trapped in the foreland. However, carbonates developed in the Vic Basin during the late under-filled stage, and were affected by siliciclastics derived from the advancing orogenic wedge and passive margin (b). Carbonates developed in association with proximal fan delta sediments in the Kasaba Basin, Turkey (c).

The relationship between fault movements in the Catalan Coastal Range and sedimentation is well documented in the Sant Llorenç del Munt fan (Monstad 2000). However, there is no evidence to suggest sedimentation at Calders was fault-controlled. Characteristics of fault-influenced sedimentary successions such as over-steepened bedding, deposition from mass flows and soft-sediment deformation (cf. Lonergan and Schreiber 1993) are not evident at Calders. Additionally, although a wide variety of carbonate facies formed, there was little variation in the relative depths of deposition (see **Chapter 3**). Carbonate-siliciclastic cyclicity at Calders is attributed predominantly to intrinsic changes in the marine system and is discussed fully in **Section 7.2.3**. On the timescale represented by the Calders depositional succession, tectonism was important in providing an elevated hinterland and siliciclastic sediment source, although accommodation space was created through subsidence related to emplacement of the orogenic wedge to the north (Taberner *et al.* 1999).

Carbonate development in the Catalan sector of the Pyrenean foreland also occurred in the northern Vic sub-Basin during the Middle Eocene (Taberner 1982; Taberner and Bosence 1985, 1995; Alvarez *et al.* 1994, Franquès 1998). The northern sub-basin received the erosional detritus derived from the advancing Pyrenean thrust sheets and led to the deposition of the Sant Martí Xic Formation (Taberner 1982, Taberner *et al.* 1999) (**Figure 7.3b**). Carbonate bioherms and fringing reef systems developed on temporarily abandoned delta lobes (Taberner and Bosence 1985, Alvarez *et al.* 1994, Franquès 1998). Sedimentation and consequent reef development was not fault controlled, although the deltaic sediments succeeding the La Trona reef complex described by Alvarez *et al.* (1995) and Franquès-Faixa (1998) are highly deformed (pers. obs.) (**Figure 7.4**).

Miocene carbonate reef systems of the Kasaba Basin, southwestern Turkey, developed on fan delta systems in close proximity to rapidly advancing thrust sheets (Hayward *et al.* 1996). The Kasaba Basin evolved into a foreland system in the Miocene through subsidence induced by the loading of the foreland by the southerly migrating Lycian Nappes (Hayward *et al.* 1996) (**Figure 7.3c**). The foreland progressively filled with a shallowing upward succession of submarine fan to fan-delta siliciclastics. Sedimentation in marginal marine areas was largely fault-

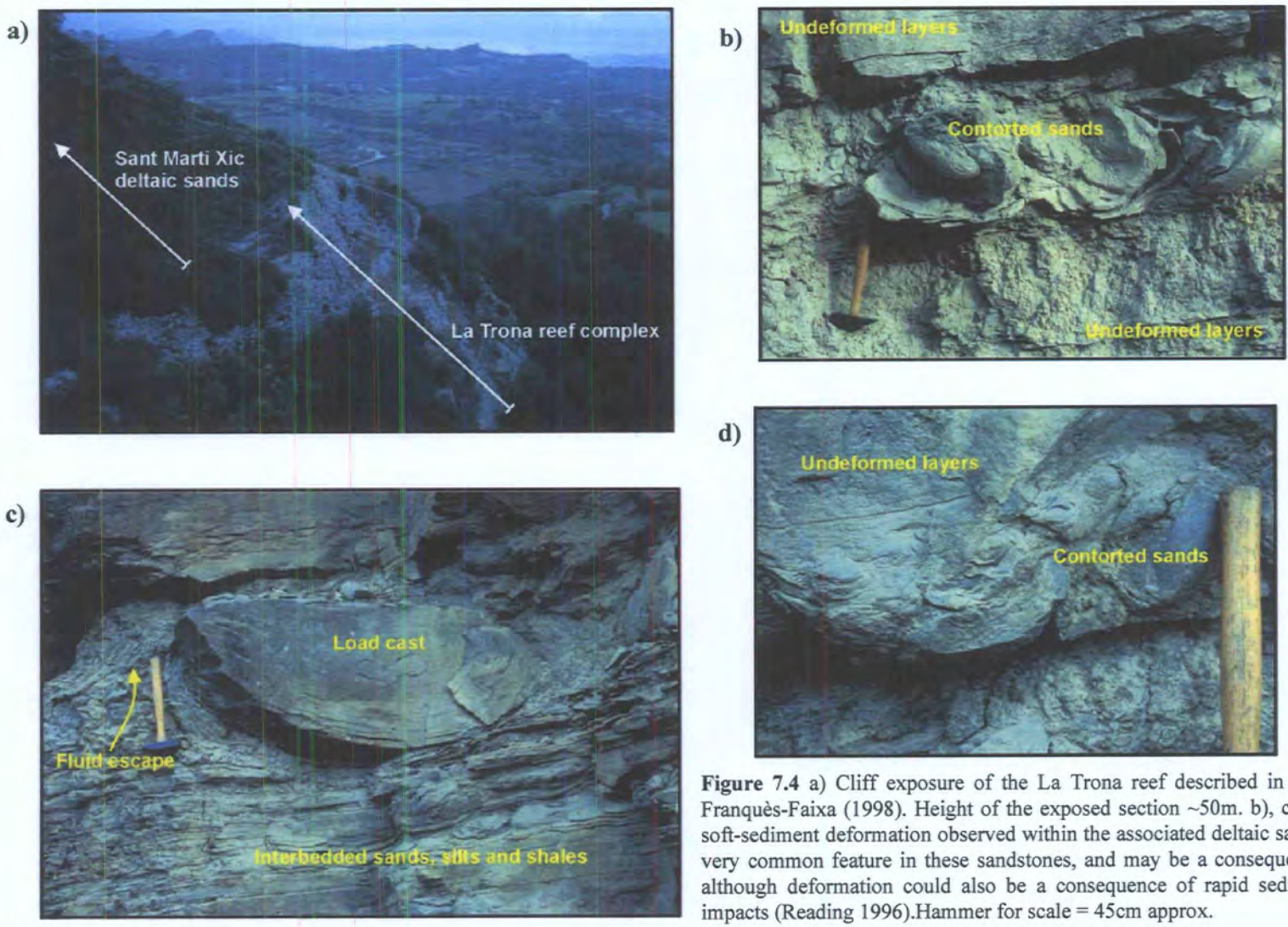


Figure 7.4 a) Cliff exposure of the La Trona reef described in Alvarez *et al.* (1994) and Franquès-Faixa (1998). Height of the exposed section ~50m. b), c) and d) illustrate styles of soft-sediment deformation observed within the associated deltaic sandstones. Deformation is a very common feature in these sandstones, and may be a consequence of regional tectonism, although deformation could also be a consequence of rapid sedimentation rates and wave impacts (Reading 1996).Hammer for scale = 45cm approx.

controlled, and sediments had short transport pathways. Importantly, rapid subsidence associated with the regional emplacement of the Lycian Nappes and extensive migration of the basin depocentre resulted in high sedimentation rates across the delta fans (Hayward *et al.* 1996). The scattered development of reefs within the siliciclastic sequence indicates conditions for growth were only satisfied sporadically, possibly as the locus of sedimentation switched during times of tectonic quiescence.

The Miocene reef systems of the Kasaba Basin, that developed in a foreland basin setting proximal to the orogenic wedge are in many ways similar to those of the Altorreal study area and the Miocene to Recent Red Sea reefs i.e. coarse grainsize, proximal location of reefs relative to the sediment source, sporadic reef development etc. However, the Red Sea reefs and the Altorreal succession were deposited in a very different tectonic setting. Subsidence and active progradation of fans in the Kasaba Basin, as a consequence of the proximal foreland location, was far more extensive (Hayward *et al.* 1996). The success of reefs in this location is thought to be related to the arid climate and the periodicity of siliciclastic sediment input (**Section 7.2.1.2**).

7.2.2.3 Foldbelts

Reefs associated with Miocene fold belts of the Mediterranean region developed on submerged, thrust-generated sea floor bathymetric highs. Tortonian to Early Messinian examples are preserved in Sicily and Tuscany (Pedley 1988 and references within). Extensive reef development in Orania, Algeria, also occurred on submarine Tortonian-aged folds during a transgressive phase in the Messinian (Saint Martin 1996). Reefs closely follow pre-existing topography with small bioherms and laterally extensive linear fringing reefs, preserved in the Traras, Tessala and Beni Chougrane Mountains (Saint Martin 1996). Carbonate development on active fold belt systems is largely inhibited by tectonically induced vast gravity slides and the development of siliciclastic sediment prisms flanking emergent orogenic zones. Colonisation by carbonate producers was inhibited largely by high sedimentation rates (Pedley 1996). Reef development in this tectonic setting would be largely restricted to the highest submerged bathymetric points away from debris flows.

7.2.3 Autogenic influences

Processes intrinsic to the depositional setting may be responsible for changes in sedimentation patterns (Tucker and Wright 1990, Reading and Levell 1996). These processes are autogenic and contrast with the allogenic processes described in **Sections 7.2.1** and **7.2.2.** Autogenic process may be an important influence on sedimentation occurring on timescales 10^3 to 10^5 years. In carbonate systems, a carbonate platform will prograde when carbonate production exceeds the rate at which accommodation space is being created, producing shallowing upward cycles over time (Tucker and Wright 1990). In a siliciclastic delta system, distributaries will avulse as steeper routes to the sea become available. Within a dynamic sedimentary environment, these autogenic changes are inevitable, although the timing may be governed by an unusual event such as a high magnitude flash flood or seismic shock (Reading and Levell 1996).

Repetitive depositional sequences are observed in both the Calders (**Figure 3.9**) and Altorreal (**Figure 5.22**). Autogenic processes such as switching of siliciclastic depocentres were important control on patterns of sedimentation (**Figure 7.5**). The Calders succession is part of the late Bartonian Marine Sequence and represents the regressive part of the second Bartonian cycle (see **Chapter 2**). Six repetitive cycles of varying thickness and completeness are identified at Calders. A typical cycle, from bottom to top, is composed of: 1) cross-stratified siliciclastic-dominated unit, often topped with a lenticular conglomeratic channel-fill, 2) a unit rich in larger benthic foraminifera with a variable clay to pebble-grade siliciclastic component and 3) *in situ* and reworked carbonate units deposited within a variety of reef environments. Slight shallowing upward trends are inferred for each cycle, but there are no indicators of significant bathymetric changes both within and between cycles (**Section 3.4.6**). However, each cycle becomes more carbonate rich vertically, indicating changes towards environmental conditions more favourable for carbonate-producing organisms (i.e. water clarity, relative depth of the photic zone, substrate stability, nutrient levels etc). Each cycle represents the reduction/cessation of siliciclastic input and colonisation of an abandoned siliciclastic substrate, initially by larger benthic foraminifera and coralline algae, and culminating with the development of a coralgal reefs (see **Section 3.4.6**). The composition of siliciclastics within each cycle does not change (i.e. calcarenitic to litharenitic sands with a variable clay

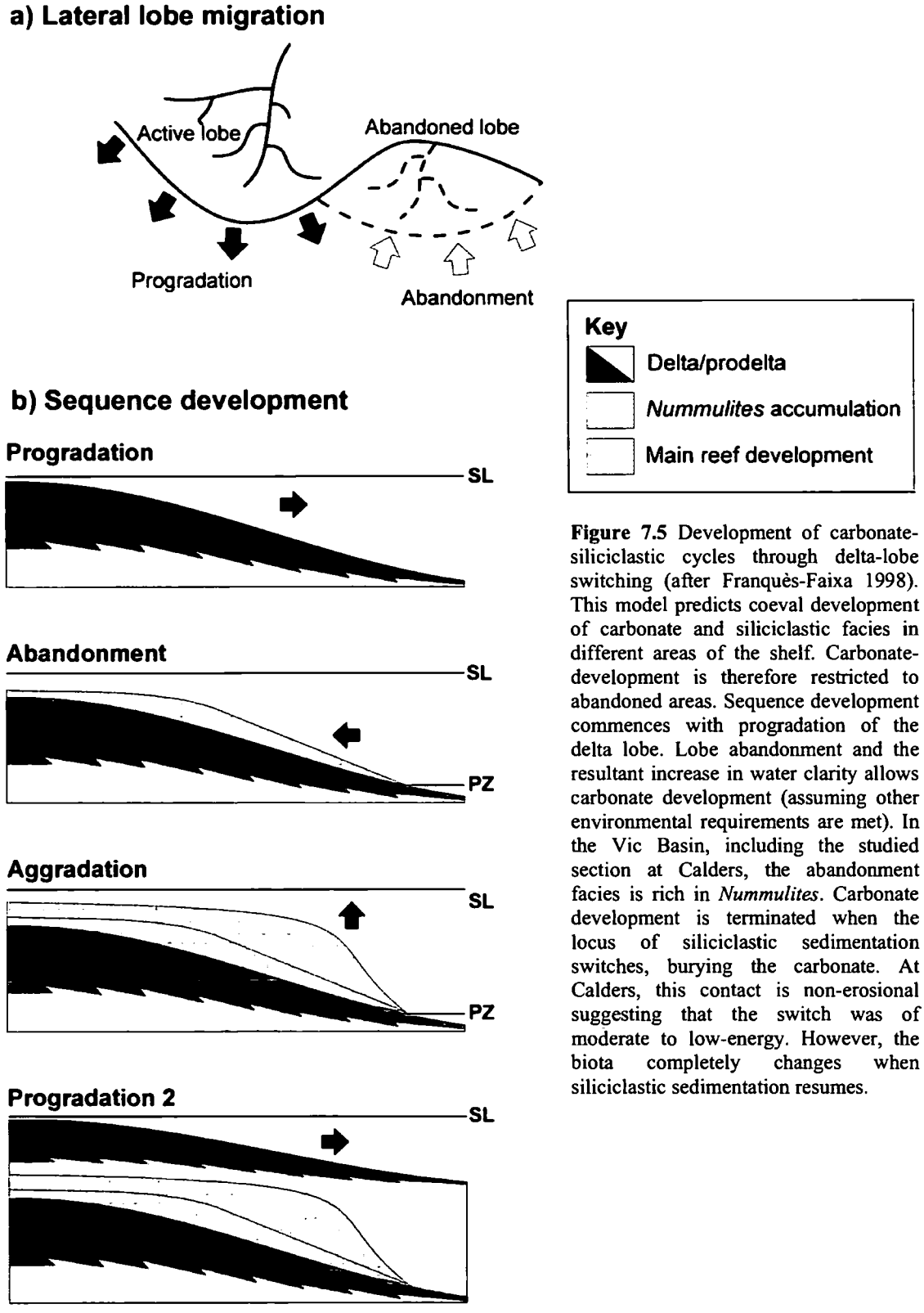


Figure 7.5 Development of carbonate-siliciclastic cycles through delta-lobe switching (after Franquès-Faixa 1998). This model predicts coeval development of carbonate and siliciclastic facies in different areas of the shelf. Carbonate-development is therefore restricted to abandoned areas. Sequence development commences with progradation of the delta lobe. Lobe abandonment and the resultant increase in water clarity allows carbonate development (assuming other environmental requirements are met). In the Vic Basin, including the studied section at Calders, the abandonment facies is rich in *Nummulites*. Carbonate development is terminated when the locus of siliciclastic sedimentation switches, burying the carbonate. At Calders, this contact is non-erosional suggesting that the switch was of moderate to low-energy. However, the biota completely changes when siliciclastic sedimentation resumes.

component) indicating that the cessation of siliciclastic input was temporary and the siliciclastics within each cycle have the same provenance. The upper and lower contacts between cycles are conformable with no evidence of erosion and non-deposition thus no catastrophic event was responsible for these cycles.

The units rich in foraminifera are dominated by *Nummulites bedai,* indicating deposition of the Calders succession within Shallow Benthic Zone 18 (SBZ 18). The duration of SBZ 18 is approximately 1.4 Ma (Serra-Kiel *et al.* 1998). Assuming sedimentation continued throughout this period in the Calders area, each depositional cycle has a mean duration in the order of 10^5 yr i.e. within the timescale of autocyclicity. This is likely to be an exaggeration, as episodes of non-deposition and erosion, together with post-depositional compaction and dissolution, will result in reduced sedimentary thickness and missing cycles. In conclusion, abandonment of the siliciclastic substrate and the development of shallow water carbonates is attributed to autogenic processes. Santisteban and Taberner (1988), Alvarez *et al.* (1994) and Franqués-Faixa (1996), attribute the development of carbonate-siliciclastic cycles in the Vic Basin to channel avulsion episodes and abandonment of delta lobes (**Figure 7.5**). It was not possible to confirm a deltaic origin for siliciclastic sediments in the Calders study area, although strong basinward currents are inferred from the development of metre-scale cross-stratified units.

Carbonate-siliciclastic cycles are defined within the succession at Altorreal in the Fortuna Basin. A typical cycle, from bottom to top, comprises: 1) cross-stratified, coarse-grained lithoclastic units with a clear progradational geometry, 2) a pebbly molluscan detrital packstone and 3) *in situ* carbonates, composed principally of corals and stromatolite. There are no significant bathymetric variations within each cycle, and, as at Calders, cycles are not inferred to shallow upward. However, each cycle indicates vertical (and lateral) changes toward conditions more favourable to photoautotrophs including increased water clarity, substrate stability and lower nutrient levels. Each cycle represents the cessation of siliciclastic input and colonisation of a temporarily inactive siliciclastic substrate, initially by molluscs and coralline algae, and culminating with the development of coral-microbial carbonate ramp-type system (see **Chapter 5**).

In contrast to cycle boundaries at Calders, there is evidence of erosion and non-deposition both within and between cycles. Initially, stromatolitic carbonates

developed in up-slope/proximal areas synchronous with hardground development in downslope/distal areas (**Section 5.2.3.1**). A prolonged period of non-deposition is indicated by hardground development. Early cementation is encouraged by the high depositional porosity and the marginal marine locality (with an influx of freshwater probably as run-off and groundwater) (**Section 5.3**). The contact between cycles is often erosional, with depth of erosion up to 1.5 metres, and erosional cavities are filled with lithoclastics (**Figure 5.22**). Karstification of some of the carbonate bodies is inferred. The dearth of erosional reef detritus within lithoclastic units indicates that erosion of the carbonate bodies was not through shifting fan delta distributaries and siliciclastic input. Erosion of some carbonate bodies is thought to be related to local uplift and subaerial exposure rather than re-activation of the delta lobe as during the late Miocene the Fortuna Basin was dominated by a transpressional tectonic regime with the generation of uplifted fault block terranes (Garcés *et al.* 2001). Eroded carbonate bodies were subsequently buried by siliciclastics, preventing immediate recolonisation. Carbonate development accompanied later episodes of delta lobe abandonment.

In summary, autogenic switching of delta distributaries and abandonment of siliciclastic substrates provides temporary sites for carbonate development, assuming other environmental conditions are acceptable for carbonate-producing photoautotrophs. Carbonate production ceases abruptly when the siliciclastic supply resumes.

7.2.4 Palaeogeographic setting and shelf morphology

Within siliciclastic-dominated environments, carbonate development is a transient condition and the short (geological) time spans over which carbonate accumulates tends to be strongly controlled by processes pertaining to the depositional environment such as the basin setting, the presence of pre-existing bathymetric highs and the hydrodynamic regime.

7.2.4.1 Basin setting and energy regime

The Calders and the Altorreal reef systems developed in comparable palaeogeographic situations in that deposition occurred in semi-isolated (but fully marine) waters (**Chapters 2 and 4**). Carbonate provinces may develop in

environments with a wide range of contemporaneous energy regimes. The incident energy on a carbonate province can affect the dominant organisms, formation of sedimentary structures, sediment grainsize and sediment sorting. The energy regime and the effects on biota development in each of the study areas are discussed below.

7.2.4.1.1 Calders energy regime

In the late Middle Eocene, the Vic Basin was situated in an oceanic inlet, open to the west with a narrow strait leading to the Atlantic Ocean. The basin was subject to moderate longshore currents (Taberner pers. comm. 2002) although it is unlikely that there was sufficient fetch across the Vic Basin to generate major storm waves.

Sediments in the Calders area were deposited in a range of energy regimes (**Chapter 3**). The highest-energy deposits are grainstones and rudstones that formed as foralgal shoals and laterally restricted coralgal biostromes respectively. Low-energy deposits are wackestones and packstones that formed in a range of environments including protected inner shelf, fore-shoal and back-shoal (**Chapter 3**). It is inferred from the absence of laterally extensive erosion surfaces and coarse, angular coral rubble that grainstone and rudstone carbonate facies are a result of *in situ* winnowing in shallow shoal type environments, not high-energy storm re-working (**Section 3.3.1**). The dominant energy regime during times of carbonate production was moderate to lower energy, with limited modification of carbonate facies distribution and architecture by waves. Biota present include a mixture of sessile (i.e. corals, articulated coralline algae, encrusting algae and foraminifera) and free-living forms (large benthic foraminifera, rhodoliths) indicating a stable substrate but energies high enough to turn rhodoliths (see **Chapter 3**). Fine-grained micrite and porcellaneous foraminifera-dominated facies formed within protected, low-energy inner shelf environments (represented by the Sant Amanc section, **Section 3.5**). The environment of deposition is inferred to be lagoonal or back-reef, and the development of small (< 1 m) patch reefs are indicative of stenohaline conditions. While there is no indication of storm reworking at this locality, modification of small patch reefs by currents is inferred from the elongate nature of carbonate bodies (**Section 3.5.1**).

Siliciclastic and mixed carbonate-siliciclastic facies formed in areas of unstable substrate (**Sections 3.3.2, 3.3.3** and **3.3.4**). The arrangement of larger benthic

foraminifera tests within the mixed carbonate-siliciclastic facies is indicative of accumulation on unstable substrates under variable energy conditions (**Section 3.4.2**). Only organisms with a degree of mobility and a tolerance to higher sedimentation rates (i.e. *Nummulites,* rhodoliths) could survive in this setting (**Section 6.4**).

Siliciclastic-dominated facies formed in a high-energy regime, with highly unstable substrates and a low faunal diversity (**Section 3.4.1**). Siliciclastic facies are arranged as metre-scale clinoforms (**Figures 3.38a** and **3.38b**), progradation of which would have required the current-mobilisation of grains. Additionally, swaley cross-statification is evident (**Figure 3.39a**) indicative of storm reworking (Leckie and Walker 1982). High-energy conditions resulted in the development and maintenance of a mobile substrate, inhibiting the establishment of an attached benthic community (**Section 3.4.1**). Carbonate producing organisms associated with these substrates include echinoids, molluscs, larger benthic foraminifera and rare rhodoliths. Where sea grass shoals are inferred to have developed, a secondary epiphytic community dominated by *Gypsina* developed (**Section 3.3.3.1**).

In summary, although Calders was not affected by high-magnitude storm events, moderate to high-energy conditions accompanying siliciclastic input yielded unstable substrates inhibiting colonisation by sessile benthic organisms.

7.2.4.1.2 Altorreal energy regime

In the Miocene, the Fortuna Basin was situated within the Betic Strait, a corridor connecting the Atlantic and Mediterranean, which remained open until the latest Tortonian (Sanz de Galdeano and Vera 1992, Garcés *et al.* 2001). As a consequence of regional uplift and uplift of faulted basement blocks, the Betic Strait was a highly sinuous route by the late Tortonian (**Figure 4.4**). It is envisaged that there was insufficient fetch to generate large storm waves thus longshore and local currents would have been relatively important.

Silty marls of the Fortuna Basin were deposited under a low energy regime (**Section 5.2.2.1**). The Altorreal system was characterised by the intermittent, high-energy input of siliciclastics within the shallow subtidal zone, with carbonates developing during quiescent periods (**Section 5.3**). Siliciclastic facies, of poorly sorted lithoclastic units, demonstrate progradational morphologies (**Figure 5.2**).

Basinward progradation of coarse lithoclastic sediments occurred through a combination of cohesive and surging mass flow events (**Section 5.2.1.1**). Substrate instability largely inhibited colonisation by sessile benthic organisms although local colonisation occurred when pauses between mass flow events were long enough to allow the settling of coral larvae (**Figure 5.7a**). The dominant fauna within this unstable environment were fast-growing large encrusting oysters (**Figure 5.7b**), often preserved fragmented in the matrix. More significant colonisation only occurred when siliciclastic input ceased and a stable hardground developed (**Section 5.3.1.3.1**). Individual reef bodies at Altorreal were not affected by storms, and carbonate developed under low-energy conditions. There are no major erosional surfaces within carbonate bodies, and corals are preserved largely *in situ*.

In summary unstable substrates, as a consequence of high-energy siliciclastic input, largely inhibited colonisation by sessile benthic organisms. Colonisation by corals in shallow water accompanied hardground formation in water depths of a few tens of metres.

7.2.4.1.3 Storm-influenced carbonates

Modern biogenic warm-water carbonates are often affected by short-lived, high magnitude environmental perturbations, such as storms and hurricanes (Geister 1983, Blanchon *et al.* 1997). The suspended load of erosional detritus and siliciclastics within hurricane waves may be transported many kilometres across a shelf (Blanchon *et al.* 1997). It is postulated that material will remain in suspension a considerable time after the storm event, effectively reducing light penetration and limiting the ability of corals that have survived to photosynthesize. Additionally, due to the energetic nature of hurricane waves, material in suspension is relatively coarse-grained and may cause damage to coral tissues through abrasion (**Section 6.1.1**). However, the low frequency of very high magnitude storm events allows time for coral recovery. The impact of these perturbations may be limited (or exaggerated) depending upon the basin setting i.e. whether the reef system is situated within an open setting or a protected setting.

It has been established that the carbonates at Calders and Altorreal were not affected by major storm events, possibly due to their isolation from large tracts of open ocean. There are abundant examples of reef systems, in both the Recent and

geological past, which have undergone modification by high-magnitude storm-generated waves.

Triassic reefs of Adnet, Austria, developed under the influence of high-magnitude storm events (Bernecker *et al.* 1999). A storm influence is evident through rapid vertical and lateral facies changes and deep erosion surfaces. A major impact of storms was the fragmentation of *in situ* coral colonies and the introduction of large amounts of siliciclastic sediment into the reef system, modifying coral abundance and distribution (Bernecker *et al.* 1999). Bernecker *et al.* (1999) also noticed a tendency towards the development of low-relief facies bodies.

The Ordovician to Early Silurian reefs of the Chicotte Formation, Anticosti Island, developed on a storm influenced carbonate platform and were subjected to non-monsoonal tropical cyclones (Wilde 1991, Brunton and Copper 1994, Long 1996). The reef was situated approximately 10° to 20°S in the subtropical belt (Long 1996). The carbonate platform was exposed with strong through currents and intense storm and wave action, leading to an overall high-energy shelf regime (Sami and Desrochers 1992, Brunton and Copper 1994). Small patch reefs and bioherms composed of tabulate corals, stromatoporoids and bryozoa were situated within vast crinoid meadows (Brunton and Copper 1994). Terrigenous influx was not important at Anticosti Island, even during storm events. However, reef demise was a result of burial by storm-reworked crinoid ossicles preserved as tempestites and megaripples (Brunton and Copper 1994). Termination, albeit temporary, of carbonate production can therefore occur through rapid burial by locally generated storm detritus. The effects of burial by skeletal detritus would be comparable to the effects of burial by siliciclastic sediments.

7.2.4.2 Shelf morphology and siliciclastic bypassing mechanisms

The presence of pre-existing bathymetric highs, and the availability of stable substrates, are important controls on the presence of carbonates in shallow marine settings where there is a siliciclastic input (Leinfelder 1997, Ferro *et al.* 1999, Cunningham *et al.* 2003). Photoautotrophic biogenic carbonate development can be initiated where the seafloor lies within the photic zone. In cases of high turbidity due to siliciclastic input or current/wave re-suspension, the seafloor may barely lie within the photic zone, severely limiting the growth, development and reproduction of

photoautotrophs. In addition, colonisation by immobile calcareous benthos, e.g. colonial corals, in areas undergoing siliciclastic sedimentation is difficult because of an unstable substrate and inability of organisms to settle in their larval stage (**Section 6.1**). Protection from siliciclastic influx may be offered by pre-existing shelf structures, of tectonic or sedimentary origin, that provide colonisation sites. Elevated sites such as uplifted basement relief (horst and graben structures, tilted fault blocks etc) and former karst relief are ideal sites as coarser grains, transported as bedload, will be trapped in bathymetric depressions (Leinfelder 1997). Bathymetric relief related to sedimentary structures and abandoned sedimentary systems may provide suitable sites for carbonate development (Cunningham *et al.* 2003). The cessation of carbonate development on the Florida Platform during the Miocene is attributed to siliciclastic input. However, prograding delta lobes provided a substratum for later rejuvenation of the carbonate platform in the Quaternary (Cunningham *et al.* 2003). A similar scenario is described from the Belize margin, where shelf-edge siliciclastic delta and slope fan wedges provided a substrate for modern barrier reef development (Ferro *et al.* 1999).

Examples of carbonate systems strongly influenced by shelf tectonism are numerous (**Section 7.2.2**). Sites of carbonate development in the Gulf of Suez in the Red Sea region have been controlled by the availability of tectonic highs since the Miocene (Purser 1987). Horst and graben structures have evolved in the gulf as a consequence of rifting. Carbonate development is largely restricted to horsts and coarse-grained siliciclastics are trapped in the intervening grabens (Purser *al.* 1987). The success of this trapping mechanism is attributed to the coarse-grained nature of sediments, which are transported as bedload (Purser *et al.* 1987). Siliciclastic grainsize in the Red Sea region is a function of the semi-arid climate and short-sediment transport pathways (see **Section 7.2.1.2**).

Near-shore siliciclastic trapping mechanisms may also be environmental i.e. estuaries, estuarine deltas, beach barrier shelter swamps and lagoons (Leinfelder 1997). In the Bahia Tapon Bay (Vieques Island, Puerto Rico), trapping of terrigenous sediment within mangroves has allowed carbonate development in adjacent embayments (D'Aluisio-Guerrieri and Davis 1988). Additionally, reef development within alluvial channels in the Red Sea is facilitated by trapping of siliciclastics behind wave-built berms (Hayward 1985).

Carbonate development at Calders and Altorreal is not thought to have been restricted to tectonic bathymetric highs and there is no evidence to suggest a nearshore environmental trapping mechanism (see **Chapters 3** and **5**). Development peaked during temporary periods of low siliciclastic input related to autogenic influences and the redirection of siliciclastics (**Section 7.2.3**).

The locus of carbonate production varies little over the period of the Marine Sequence represented by the sedimentary succession at Calders (**Section 3.4; Figures 3.43**, **3.44** and **3.46**). Carbonate production was halted through the variable effects of siliciclastic input, although when siliciclastic input ceased/decreased sufficiently to allow recolonisation by photoautotrophs, carbonate development resumed in a similar position on the shelf, producing vertically-repetitive facies cycles (**Figure 3.9**). It is suggested that this is not purely a coincidence, and that carbonate accumulations developed as minor topographic highs, possibly demonstrating only a few centimetres relief (**Chapter 3**). These minor depositional highs may have been preserved within overlying siliciclastic facies, providing locally elevated sites for future carbonate development. Additionally, differential compaction of siliciclastics and carbonate sediments can produce bathymetric variations. Mechanical compaction will begin as soon as soon as there is overlying sediment (Tucker and Wright 1990). The effects of compaction will be most marked in fine-grained siliciclastic and carbonate sediments compared to grainy sediments (**Figure 7.6**). It is postulated early cementation within skeletal and grainy facies such as the coral mixstone (**Section 3.3.1.1**) and foralgal grainstone (**Section 3.3.1.7**) would make them less susceptible to compaction than adjacent siliciclastic and mud-rich carbonate facies. With progressive burial, bathymetric highs will be accentuated above carbonate deposits (**Figure 7.6**). This process is thought to have been less important at Altorreal due to the coarse-grained nature of siliciclastics and the development of carbonates within bathymetric lows as laterally restricted carbonate ramp-type deposits (**Section 5.3.2**).

Reef initiation in an area of active siliciclastic input off the coast of Bahia, Brazil, was facilitated by the presence of antecedent reefal bathymetric highs (Leão and Ginsburg 1997). Coral colonisation of the inner shelf of the Great Barrier Reef has been restricted through the presence of a post-glacial terrigenous muddy wedge (Belperio 1983, Larcombe and Woolfe 1999). However, exposure of late Pleistocene to early Holocene cobbles and conglomerates has provided stable substrates in inner-

shelf areas where water depth < 20 m (Woolfe and Larcombe 1998, Larcombe *et al.* 2002). Reefs are absent in the inner shelf at depth greater than 20-40 m as a consequence of high turbidity (Woolfe and Larcombe 1999).

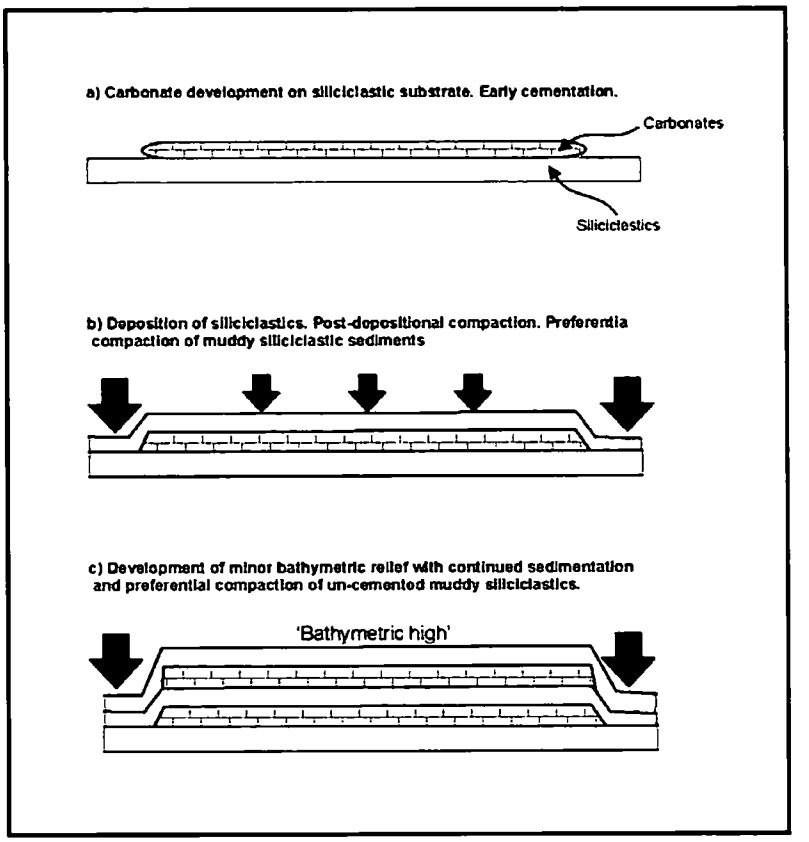


Figure 7.6 Stylised development of minor bathymetric relief as a consequence of the differential compaction of cemented carbonates and muddy, uncemented siliciclastics. The bathymetry generated may only be a few centimetres above the sea floor, but this can raise the substrate into the photic zone in turbid waters.

During sea level lowstand, siliciclastics may completely bypass the inner shelf. The Miocene Ziglag and Pattish Formations, deposited within the eastern Mediterranean region, represent extensive fringing reef development in association with coarse-grained siliciclastics (Buchbinder 1996). The basin margin was characterised by a local continental drainage system feeding coarse siliciclastic material to the coast mainly as gravity flows. Siliciclastics bypassed the inner shelf, as gravity flows were channelled into submerged canyons depositing debris sheets that evolved into submarine fans (Buchbinder 1996). Conditions in the inner shelf were therefore ideal for reef development, evident through the development of laterally extensive fringing reef systems (Buchbinder 1996).

7.2.4.3 Hydrodynamic trapping and diverting mechanisms

Sites of carbonate development are often away from sites of significant siliciclastic input such as temporarily abandoned delta lobes (**Section 7.2.3**). Suitable substrates can develop through the action of local and regional currents that manipulate suspended sediment transport pathways. This may be one of the most important controls on reef development in siliciclastic settings over short time periods. Wind-generated longshore currents that develop offshore may trap particulate material forming a turbid coastal boundary layer (Leinfelder 1997, Woolfe *et al.* 2000). Trapping of sediment in these boundary layers limits the shelfward migration of turbid waters by deflecting turbid water plumes to an orientation virtually parallel to the coastline.

Coral and *Halimeda* patch reef systems have developed in delta front environments in close association with siliciclastic sediments of the Mahakam Delta, Borneo since the Miocene, despite a massive sediment input to the coast from the Mahakam River of 8×10^6 m^3 $year^{-1}$ (Allen *et al.* 1976, Wilson and Lokier 2002). Recent carbonate-producing areas are concentrated around the northern parts of the delta. This distribution is because the southern lobe is currently more active than the northern lobe. In addition, the southward flowing Indonesian Throughflow Current results in relatively less turbid waters to the north (Roberts and Sydow 1996, Wilson and Lokier 2002).

The presence of longshore current systems and restriction of suspended material to a brackish coastal boundary layer, has allowed the development of a large offshore carbonate province in the Caribbean, off the coast of Nicaragua (Roberts 1987, Murray *et al.* 1988, Leinfelder 1997). The humid setting with high precipitation rates, in conjunction with significant hinterland relief and high continental runoff, has resulted in constant high siliciclastic input to the coastline. Freshwater discharge is approximately 1.45×10^{11} m^3 yr^{-1}, supplying an estimated sediment input to the coast of 24 to 32×10^6 mt yr^{-1} (Murray *et al.* 1988). The brackish, turbid coastal boundary layer is very wide, ranging from 20 to 30 km limiting extensive coral reef development to the shelf edge where waters are clear (Murray *et al.* 1988).

High-energy inner-shelf perturbations can also favourably manipulate terrigenous sediment pathways. The Ordovician to Silurian carbonate ramp system of

Anticosti Island, Canada, developed in a foreland basin setting on the Laurentian margin, and was subjected to severe tropical cyclones during a marked warm season (Wilde 1991, Long 1996). There was a moderate siliciclastic input into the carbonate ramp via delta and estuarine sources. Under fair-weather conditions, fine-grained detritus may have been confined to the inner parts of the shelf by coastal boundary currents. (Long 1996). However, siliciclastics bypassed the inner shelf and were redistributed by hurricane-driven storms across the ramp producing clear-water conditions along the inner shelf (Long 1996).

7.2.4.4 Turbidity and sedimentation

Despite the myriad case studies that conclude that carbonate development is normally away from sites of direct siliciclastic input e.g. on the inactive portions of a delta system or a bathymetric high (**Sections 7.2.4.2** and **7.2.4.3**), "net sediment supply into coastal environments cannot be directly correlated with the regional presence or absence of coral carbonate provinces" (Woolfe and Larcombe 1998). Sediment supply to the inner shelf may vary temporally and spatially, often by several orders of magnitude. The sites and longevity of coral development are affected by autogenic factors such as delta lobe switching (**Section 7.2.3**), but also shorter-term environmental perturbations, in particular turbidity, related to the local and regional hydrodynamic regime. Turbidity may be generated through the re-suspension of fines by wave action and/or unidirectional currents (Larcombe and Woolfe 1999). This requires a source of fines for re-suspension.

The main phases of carbonate development at Calders and Altorreal coincided with the virtual cessation of sandy or coarse-grained siliciclastic input (see **Chapters 3 and 5**). This is related to phases of abandonment of the siliciclastic substrate (**Sections 3.4, 5.3** and **7.2.3**). However, coral-dominated carbonate facies at Calders may contain up to 38 % non-carbonate material by weight (see **Chapter 6**). The grainsize of this material ranges from clay to coarse silt, probably input gradually as fallout from suspended load (**Section 3.4**). The amount of material in suspension at Calders during times of coral development cannot be quantified directly as turbidity is a physical attribute of a system that exists at a point in time and does not necessarily leave a physical deposit. In addition, fine-grained sediment may be continually re-suspended and re-deposited, with no net sediment input or transport (cf. Roy and

Smith 1971, Woolfe and Larcombe 2000). The amount of non-carbonate material in the matrix of reef facies provides a lower limit for siliciclastic input (**Chapter 6**). Fine-grained siliciclastics within the matrix of carbonate facies are likely derived from the re-suspension of material from siliciclastic facies in the vicinity. From the laterally extensive nature of carbonates at Calders (> 2 km parallel to the palaeoshoreline), it is inferred that suspended material was transported several kilometres from adjacent siliciclastic-rich areas (**Chapter 3**). There may also have been a degree of *in situ* reworking of siliciclastics by wave action. If turbidity is to be inferred for periods of reef growth at Calders, the duration of turbid water events and the sedimentary impacts were at sub-lethal levels (cf. Larcombe *et al.* 2000). In addition, the morphological effects of turbidity on corals, e.g. the development of thin, laterally extensive platy morphotypes (see **Chapter 6**) is not apparent at Calders. The transport of suspended material is therefore thought to have been periodic, possibly related to seasonal climatic variations (**Section 7.2.1.1**).

Re-suspension of siliciclastics would not have been an important influence on carbonate development in the Altorreal area. Siliciclastic sediments range from sand to cobble grade with negligible clays and silts (**Section 5.2.1**), and would have been too coarse to be re-suspended under normal fair weather conditions (Hayward 1985, Ahmed *et al.* 1993, El Sammak *et al.* 1997). This is confirmed from the absence of siliciclastics within carbonate units (**Section 5.2.3**). However, re-working of coarse material into branching coral thickets is inferred along the margins of the reef in shallow water areas (**Section 5.2.3.3**). This is not thought to have been detrimental to corals, but coarse siliciclastics probably supported colonies in high-energy conditions (**Section 5.2.3.3**).

The re-suspension and transport of fine-grained material can lead to reefs being affected by siliciclastic input some distance from the sediment source. The local and regional hydrodynamic regime plays an important role on the distribution of sediments and turbidity on the inner shelf of the Great Barrier Reef (Belperio *et al.* 1983, Belperio 1988). The inner shelf is strongly influenced by episodic wave-induced re-suspension of fine material (Larcombe and Woolfe 1999). Discrete turbidity events are 60 to 100 hours in duration (Larcombe *et al.* 2002). Particulate sediments in Halifax Bay, a site of turbid zone reef growth in the central Great Barrier Reef, are derived from the Burdekin River Delta situated 200 km to the south (Woolfe

and Larcombe 1998, Larcombe *et al.* 2001). The Burdekin River supplies 2.7 to 8.5 Mt of terrigenous sediment to the inner shelf annually (Moss *et al.* 1993, Neil and Yu, 1995). Particulate sediments are transported northward (parallel to the shoreline) by the SE trade winds and tidal currents. During these high turbidity episodes, insufficient sediment is deposited on the corals to cause smothering as sediment is continually re-suspended and transported alongshore by wave and wind-driven currents (Larcombe and Woolfe 1999).

Turbidity around fringing reefs off the coast of Phuket, Thailand, is attributable to the re-suspension of mud-grade material through tidal currents and wave action (Tudhope and Scoffin 1994). Transport and re-suspension has resulted in a gradient in water clarity from inshore to offshore areas (Tudhope and Scoffin 1994). Suspended siliciclastic sediments affect the Albrolhos Reefs, off the coast of Bahia, Brazil, although the reefs are situated >200 km from the siliciclastic source (Leão and Ginsburg 1997). Under normal fair weather conditions, transport of coarse siliciclastics is limited to the inner shelf and muds flocculate and tend to be deposited in near-shore areas (Leão and Ginsburg 1997). Strong longshore currents affect the coast during winter storms and reefs are subjected to high turbidity (Leão and Ginsburg 1997).

7.2.4.5 Size and morphology of reefs in siliciclastic settings

The morphology of carbonate units is dependent upon a variety of factors including the type of carbonate producers, the presence/absence of antecedent topography, depth range of carbonate producing biota, relative sea-level changes and the period of carbonate development (Wilson and Lokier 2002). Carbonates in mixed carbonate-siliciclastic sedimentary successions demonstrate a variety of morphologies and are extensive laterally from a few metres to many tens of kilometres.

The carbonate units at Calders and Altorreal reflect the depositional morphology of the siliciclastic substrate. Laterally extensive (> 2 km) carbonates and low-angle clinoforms reflect the wide, low-angle morphology of the siliciclastic shelf at Calders (**Section 3.4**). Coral development was localised with patchy *in situ* framework development (**Chapter 3**). Most carbonate facies are detrital, consisting of coral fragments floating within a mixed carbonate-siliciclastic matrix, and form low-angle clinoforms and tabular bodies with no evidence of topographic relief (**Section**

3.3.1). In comparison, the relatively narrow, steep shelf morphology at Altorreal has produced laterally restricted (< 100 m down-dip) sigmoidal carbonate units (**Section 5.2.3, Figures 5.14** and **5.20**). Coral development was dense forming a framework with steep slopes (**Section 5.2.3**).

Carbonates within semi-arid fan delta systems develop as large fringing reefs at the shelf slope break forming clinoform-shaped deposits, smaller patch reefs and accumulations within abandoned fan-delta channels (Santisteban and Taberner 1988, Braga *et al.* 1990, 1995, Mankiewicz 1995). During the late Miocene, fringing reef systems several kilometres in length developed on the break of slope of large fan-delta systems in the Fortuna Basin (Santisteban and Taberner 1988, Poisson and Lukowski 1990, Mankiewicz 1995), the Lorca Basin (Wrobel and Michelzik 1999) and the Granada Basin (Braga *et al.* 1990, Braga and Martin 1996). Within the Red Sea region, Miocene to Recent reefs associated with fan delta siliciclastics range from small reefs (100 to 350 m across) that developed on structural highs and within abandoned channels, and as kilometre-scale fringing reefs seaward of alluvial fans (Hayward 1985, Purser *et al.* 1987). Siliciclastics have limited influence on the type and morphology of biota that develop in semi-arid settings because sediment input is intermittent and does not typically accompany carbonate development (**Section 7.2.1.2**). Reef systems that develop in these settings tend to be high relief with vertical walls and overhangs.

In tropical, humid environments carbonates may experience near-continuous siliciclastic input (**Section 7.2.1.1**) that exerts control on carbonate development through influencing the biota present, biota distribution and deposit morphology (Wilson and Lokier 2002). Distinct differences in reef size, morphology and environments between turbid water and non-turbid water reefs are well demonstrated on the Panwa Peninsula, Thailand. Reefs in muddy environments develop wide (>300 m) inter-tidal reef flats (where coral cover is highly variable), a narrow linear reef front of large living coral colonies and a mud-dominated fore-reef (Tudhope and Scoffin 1994). In clearer water areas, which coincidentally are more hydraulically exposed, reefs develop as more isolated clumps of massive corals and branching coral thickets situated amongst sands and gravels (Tudhope and Scoffin 1994). The clear-water reefs are lacking inter-tidal reef flat areas, and the reef front is irregular (Tudhope and Scoffin 1994). Carbonates that developed in turbid waters of the

Mahakam Delta, Borneo, are up to 4 km across with very low relief with no evidence of framework development (Wilson and Lokier 2002, Wilson (*in press*)).

In summary, it is inferred that the affect of siliciclastic input on carbonate morphology is quite different in semi-arid and more humid settings experiencing siliciclastic input. The siliciclastic morphology and the presence of bathymetric highs is a fundamental control on sites and the extent of carbonate development, and basic carbonate morphology. In low-energy areas of constant, fine-grained siliciclastic input, carbonates tend to be low relief with limited *in situ* framework development. In semi-arid areas, carbonate development is not continuously affected by siliciclastic input and deposits tend to demonstrate higher topographic relief and evidence for framework development.

7.3 Sedimentary and biological components of reefs in siliciclastic settings

Organism biology is a major control on carbonate production. The biological components of a reef system in turn reflect water temperature, chemistry, pH and salinity, and to a certain extent physical parameters such as water depth, energy level and the nature of the substrate (James and Kendall 1992). Reef systems that develop in siliciclastic-influenced settings often show differences in the biota, in terms of diversity and morphology, when compared to communities of clear water reefs.

Through acid digestion analysis of carbonate units, it has been shown that there was a significant quantity of clay to silt-grade siliciclastic material in the matrix (**Chapter 6**). Slightly enhanced nutrient levels as a consequence of this siliciclastic input are inferred from the flourishing coralline algae, the high incidence of bioerosion of the *in situ* reef framework, and abundant detrital muddy matrix (Hallock 1988).

Continentalisation of the Vic Basin (and the southeastern Pyrenean foreland) occurred in the late Priabonian (Puigdefabregas and Souquet 1986). Significant evaporite development commenced in the Vic Basin during that time (the Cardona Formation) (Taberner 1982, Hendry at al. 1999, Taberner et al. 1999, Taberner *et al.* 2000). The abundance and diversity of stenohaline organisms throughout the succession demonstrate that salinity was never a control on reef development or demise at Calders (**Section 3.2**). Although there was not a connection with the

Mediterranean through this portion of the foreland basin in Bartonian time, the basin was still open to the Atlantic through a western connection (pers. comm. Taberner 2002).

As would be expected, different biotic assemblages are present in shallow-water siliciclastic-dominated environments compared with carbonate-dominated environments (**Section 3.3**). An intermediate community is present within transitional environments i.e. where there is a significant siliciclastic component within a carbonate lithology and vice versa (**Section 3.3.2**). The diversity of organisms present within reef and reef-associated environments at Calders are due in part to the diversity of habitats (**Section 3.2, Table 3.2**). Hermatypic corals, coralline algae and large encrusting foraminifera dominated shallow carbonate environments (**Section 3.3.1**). Siliciclastic environments had a relatively impoverished fauna, and were dominated by infaunal echinoids, *Cerithium* gastropods, rhodoliths and large benthic foraminifera (**Sections 3.3.4** and **3.3.4**). Transitional environments at Calders are represented by current-reworked tabular accumulations of larger benthic foraminifera. Accumulations are largely monospecific, composed of a single species of nummulite (*Nummulites bedai*) with subordinate *Discocyclina, Operculina* and *Amphistegina.* A summary of main biota in the studied sediments at Calders is presented on **Table 7.3**.

A low diversity fauna is present within both siliciclastic and carbonate deposits at Altorreal in comparison with those at Calders (**Table 7.3**). Towards the end of the Tortonian, the Fortuna Basin was becoming gradually isolated from the Mediterranean as the Betic Strait closed (**Section 4.3**). However, there is no evidence within the studied sediments to suggest enhanced salinities (**Section 5.2**).

Different biotic assemblages are present in shallow water siliciclastic-dominated environments compared with carbonate-dominated environments (**Table 5.1**). Robust oysters are the dominant organism within siliciclastic-dominated facies (**Section 5.2.1**) as a consequence of their ability to quickly colonise available substrates. A more diverse community developed with the cessation of siliciclastic input, comprising oysters, pectens, gastropods, coralline algae, serpulids and benthic foraminifera (miliolids and *Amphistegina*). In up-slope areas, laminar stromatolite development is associated with a freshwater input and possibly enhanced nutrients (**Section 5.2.2.2** and **5.3.1.2**). *In situ* corals with gastropods, bivalves, coralline algae and benthic foraminifera dominated carbonate environments. Morphological changes

Facies	Dominant Biota	
	Calders and Sant Amanc, NE Spain Late Middle Eocene (Bartonian)	**Altorreal, SE Spain Late Miocene (Tortonian-Messinian)**
Carbonate-dominated *In situ* reef environments with associated detrital facies. Siliciclastics may be present in the matrix in variable quantities	Hermatypic corals, encrusting, branching and articulated coralline algae, rhodoliths, bivalves, brachiopods, echinoids and bryozoa. High occurrence of encrustation (mainly by coralline algae and foraminifera) (**Section 3.3.1**). Often demonstrate a very diverse non-symbiont bearing foraminiferal assemblage comprising miliolids, *Gypsina, Calcarina, Haddonia, Fabiania,and Gyroidinella* (**Section 3.3.1**) Siliciclastics are locally important, and are dominated by detrital clays and silts (**Section 6.3**)	Corals (low-diversity fauna dominated by *Porites*) (**Section 5.2.3**). Microbialte as laminar 'mats' and concentric crusts on branching corals (**Section 5.2.3.3**) Lithoclasts locally present between coral branches (**Section 5.2.3.3**).
Transitional facies Increasing or decreasing carbonate production (relative to siliciclastic input) and establishment of a more sediment-tolerant community	Foraminiferal pack-grainstones dominated by mono-specific *Nummulites* with *Discocyclina*, *Operculina*, *Gypsina*, miliolds, coprolites, solitary corals and assorted skeletal fragments (**Section 3.3.2**).	Assorted skeletal fragments including oysters, pectins, miliolids, coralline algae, bryozoa, gastropods and encrusting larger foraminfera (**Section 5.2.2.2**)
Siliciclastic-dominated Inner-shelf, marginal-marine environments characterized by unstable substrates. Often demonstrates progradational sedimentary geometries	Echinoids, gastropods (*Certihium*), *Nummulites* and *Gypsina* (**Sections 3.3.3** and **3.3.4**).	Large oysters encrusting boulders with variable skeletal fragments. Rare *in situ* coral colonies (**Section 5.2.1**).

Table 7.3 Comparison of the dominant biota present in carbonate-dominated, siliciclastic-dominated and transitional facies of the studied sections at Calders and Altorreal.

of coral colonies are related to changes in water depth rather than siliciclastic input (**Section 5.3.1.3**). Coral growth is only thought to have occurred in association with siliciclastics in up-slope areas where stick-like branching corals dominate (**Section 5.2.3.3**). It is postulated that upslope areas were affected by freshwater input due to the development of stromatolitic coatings on branches (**Figure 5.18**).

A number of possibilities are suggested for reef termination at Altorreal. A freshwater input, possibly as a precursor to the reactivation of the fan delta lobe, had a detrimental effect on seawater salinity and nutrient levels. This is evident from the development of columnar stromatolites (**Sections 5.2.3.2** and **5.3.1.3**). Enhanced nutrients often hinder the development of photoautotrophs, although it has been shown that some species of coral can adapt from an autotrophic to heterotrophic mode of life (Antony 2000, Antony and Fabricus 2000). Additionally, erosion of reef bodies is observed, with siliciclastic sediments onlapping the irregular contact (**Figure 5.16**).

7.4 Synthesis: mixed carbonate-siliciclastic sequence development

An important aspect of the study of mixed shallow marine successions is the development of a depositional model of carbonate production within siliciclastic settings applicable to a diversity of successions. Excluding a volume of work published in the last century (Doyle and Roberts 1988), and a recent study by Lokier and Wilson (2002), limited attention has been paid to mixed carbonate-siliciclastic sequence development. While it is not the aim of this thesis to produce a full sequence stratigraphic model for all types of carbonate-siliciclastic succession, this section will outline the fundamentals of a potential model using observations from the two studied sections in NE and SE Spain and further examples from the literature.

7.4.1 Sequence stratigraphic models

Arguably, there is no universally accepted sequence stratigraphic model for mixed carbonate-siliciclastic successions. In the majority of examples presented in the literature, carbonates within marine mixed successions have been classified as highstand deposits. It is improbable that every biogenic carbonate accumulation within a marine siliciclastic-dominated succession should correspond to a highstand. The study of the successions at Altorreal and Calders has provided insight into

mechanisms for successful biogenic carbonate production and reef development within environments atypical of classic reef models.

Mutually exclusive depositional models have been used to describe the evolution of carbonate and siliciclastic sequences. This is because most carbonate in shallow marine environments is produced *in situ* by organisms whose diversity is typically a function of relative depth of the photic zone. It is noted by Schlager (1992) that most biogenic carbonate is produced in less than 10-20 m water depth, and that carbonate productivity is reduced rapidly below these depths to the base of the photic zone, typically around 50-100 m water depth (**Figure 7.7**). The stratigraphic response of a shallow marine carbonate system to variations in relative sea level will be governed by its growth potential and hence the productivity of carbonate producing organisms in the shallow euphotic zone (Schlager 1981, Kendall and Schlager 1981, James and Kendall 1992, Bosscher and Schlager 1992, Bosscher 1992). In contrast, siliciclastic systems are dependent upon the external sediment supply into the depositional environment that can be cut off and restarted at any water depth (Schlager 1992). The response of a carbonate system to (relative) changes in sea-level has been reviewed in detail by Schlager (1981, 1989, 1991), Kendall and Schlager (1981), Burchette and Wright (1992), Hunt and Tucker (1993), Tucker *et al.* (1993) and Handford and Loucks (1993) and thus is not reviewed here.

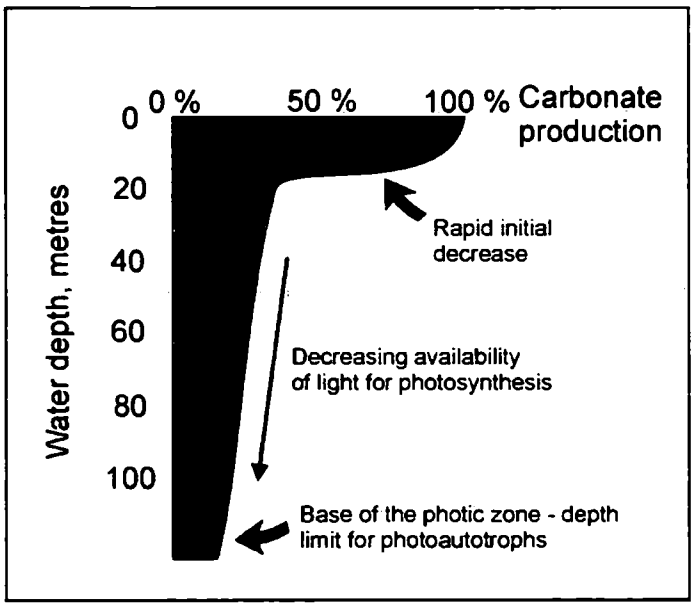


Figure 7.7 Stylised illustration of the rapid decrease in carbonate production with depth (after Schlager 1992). A similar decrease in carbonate production will also occur with an input of suspended siliciclastic sediment, which would reduce light penetration.

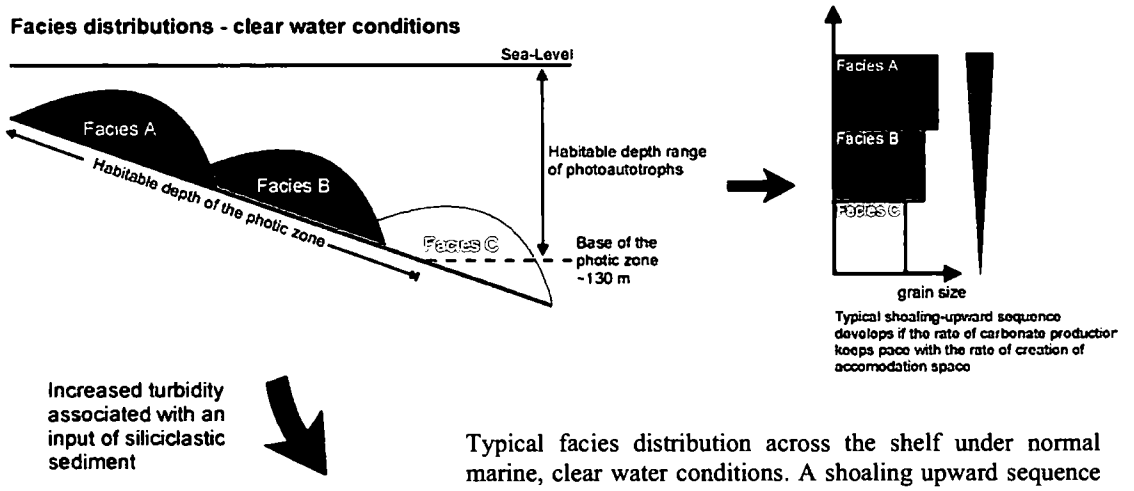
The thickness and geometry of a carbonate succession records the relationship between the rate of carbonate production to available accommodation space, eustatic sea-level change and tectonically driven subsidence (or uplift). Relative sea-level change is the sum of tectonic rates (uplift or subsidence) and rates of eustatic change. The accommodation space provided by a relative change in sea level represents the accumulation potential of a carbonate sequence (Sarg 1988). Facies distributions, thickness and geometry of the carbonate sequence are ultimately dictated by the response of the carbonate system to the creation of accommodation space (Sarg 1988).

7.4.2 Relative sea-level changes associated with siliciclastic input

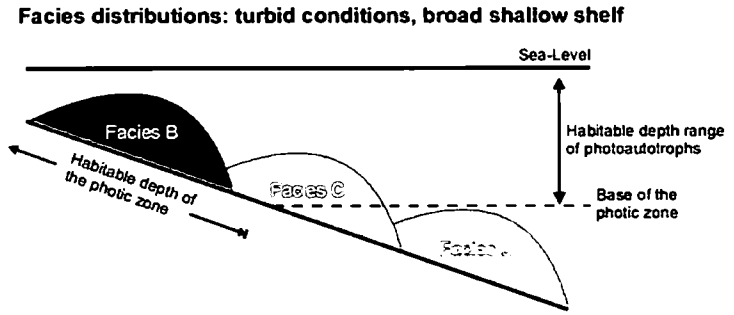
In mixed carbonate-siliciclastic successions, the conventional interpretation of depositional sequences is that they are due to reciprocal sedimentation in response to relative sea-level changes (Mack and James 1985, Buchbinder 1996, Rankey 1997). Accordingly, transgressive and highstand deposits are composed of carbonate deposits, and lowstands of siliciclastics (Ahmed *et al.* 1993, Holland 1993, Holmes and Christie-Blick 1993). The development of carbonate-siliciclastic cycles in response to glacio-eustatic sea level fluctuations is typical of Permo-Carboniferous strata (Mack and James 1985, Rankey 1997).

During a transgression, the locus of siliciclastic sedimentation moves shoreward and accommodation space becomes available allowing carbonate development within areas of the sea floor that are within the photic zone. Highstand carbonates are therefore 'pure' and show little mixing with siliciclastic facies. Upon emergence, pure carbonates are prone to erosion, dissolution and karstification (García Mondéjar and Fernández Mendiola 1993, Rankey 1997, Ferro *et al.* 1999). Highstand carbonates have been reported from the modern (Belize - Ferro *et al.* 1999, Florida –Cunningham *et al.* 2003) and ancient (Miocene, Mediterranean – Buchbinder 1996, Saint-Martin 1996; Mesozoic, Spain – García Mondéjar and Fernández Mendiola 1993; Tertiary, Spain - Luterbacher *et al.* 1997; Paleozoic, USA – Holland 1993, Rankey 1997).

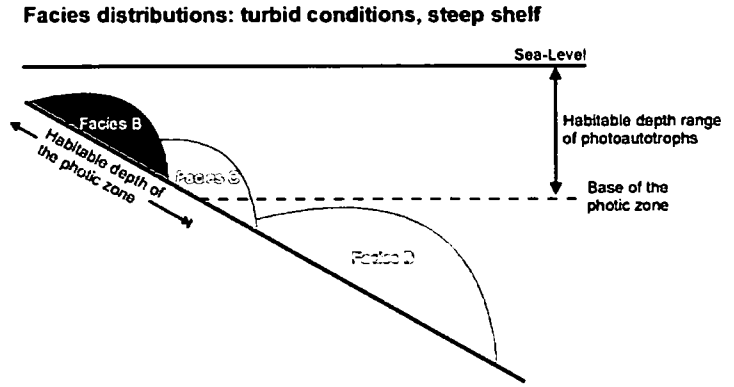
Biogenic carbonate development will only occur during a transgression if there is a decrease in siliciclastic input and improvement in water clarity to compensate for the increase in water depth **(Figure 7.8)** (Wilson, *in press*). An



Typical facies distribution across the shelf under normal marine, clear water conditions. A shoaling upward sequence will develop if the carbonate facies prograde basinward



Schematic illustration of changes to the facies distribution across the shelf after the onset of turbidity. Material in suspension will decrease light penetration, reducing the depth of the photic zone. This will reduce the habitable depth of the shelf by photoautotrophs. Consequently, facies belts will be displaced shoreward. The shallowest water facies (i.e. Facies A) may be lost altogether.



For an equivalent reduction in the depth of the photic zone on a steeper shelf, compared with a gentle slope, the habitable area for photoautotrophs will be narrower.

Figure 7.8 Schematic illustration of the postulated effects of turbidity on facies distribution on broad shallow and steep narrow shelves.

increase in water depth, or persistently turbid conditions, will reduce the habitable depth of the shelf (**Figure 7.8**). The effect of this will depend upon factors such as rate of relative sea-level rise, the amount of suspended siliciclastic material and the morphology of substrate (**Figure 7.8**). Drowning of carbonates can occur if the rate of carbonate production cannot keep pace with the rate of relative sea-level change (Schlager 1999; Wilson, *submitted*). Consequently, transgressive carbonates within carbonate-siliciclastic cycles often demonstrate deepening upward trends (Holland 1993, Rankey *et al.* 1999). Highstand shelves can still be subject to turbid conditions through the action of longshore currents (Saint-Martin 1996, Larcombe *et al.* 1999, Larcombe and Woolfe 1999). On the modern Mahakam Delta, a relative still stand in sea level during a highstand has resulted in renewed progradation of siliciclastics onto the shelf (Roberts and Sydow 1996). Highstand carbonate production is highly localised due to turbidity and siliciclastic input, with *in situ* mixing of carbonate and siliciclastic sediments (Wilson, *submitted*).

7.4.3 Carbonate-siliciclastic sequence development in the Vic Basin, NE Spain

Sediment input to the eastern Pyrenean Foreland has varied over time with periods of rapid sedimentation corresponding to allogenic processes such as tectonic uplift and erosion of the hinterland (Puigdefabregas and Souquet 1986, Burbank *et al.* 1999). Bartonian sediments of the Vic Basin were deposited during a late Middle Eocene highstand (**Figure 2.20**). On the scale of the deltaic Centelles Formation, the Bartonian-Priabonian part of which forms the focus of **Chapter 3**, sediment supply to the shallow marine siliciclastic shelf was localised due to autogenic processes such as delta lobe switching (Taberner 1983, Taberner and Santisteban 1988, Alvarez *et al.* 1994, Franquès-Faixa 1998, Taberner *et al.* 1999). Shoreward or lateral movement of the locus of siliciclastic influx resulted in an apparent, localised transgression, providing conditions that were locally favourable for carbonate development. Metre-scale carbonate-siliciclastic cycles, comparable to those described from the Calders area (**Chapter 3**), occur throughout the Middle Eocene Marine Sequence on the northern and southern margins of the Vic Basin (**Chapter 2**). Carbonate intervals are not coeval (Burbank *et al.* 1999, Taberner *et al.* 1999) and are of limited lateral extent, thus confirming that autogenic processes were important.

The thickness of carbonate cycles is variable and there are no patterns in thickness of carbonate or siliciclastic intervals within cycles (**Figure 3.9**). The diversity of depositional settings represented by carbonates is attributed to localised variations in water energy, turbidity and siliciclastic input rather than changes in water depth (**Section 3.3**). The absence of erosion surfaces, and evidence to suggest karstification or a sedimentary hiatus, indicates that adequate accommodation space was generated to allow carbonate development. As noted above, the effects of abandonment of the siliciclastic substrate are akin to an apparent transgression. However, the accumulation of 20 m of carbonates (e.g. carbonate intervals 5 and 6, **Figure 3.9**), requires a relative sea-level rise and/or subsidence. A relative sea-level rise is discounted for the reasons described above, therefore subsidence was important. The Bartonian Marine Sequence was deposited during the late under-filled stage of foreland basin development when subsidence rates are generally low (DeCelles and Giles 1996). On the southern margin of the Pyrenean Foreland (i.e. the Vic Basin), subsidence was enhanced through tectonism in the Catalan Coastal Ranges (Taberner *et al.* 1999). The carbonates at Sant Amanc, which were the final marine sediments to be deposited in the SE Pyrenean Foreland, were deposited in a significantly different setting to the sediments at Calders (**Section 3.5**). The change from oligotrophic open marine to mesotrophic partially protected is attributed to a combination of factors, but gradual basin exposure through reduced subsidence and changes in the basin configuration (i.e. basin isolation) are thought to be the most important (Travé 1992, Travé *et al.* 1996).

Preservation of the carbonate sediments at Calders is attributed to highstand siliciclastic input. Although subsidence and siliciclastic abandonment provided accommodation space for carbonate development, carbonates would have been susceptible to erosion and/or karstification during the subsequent lowstand and basin emergence (Garcia Mondejar and Fernandez Mendiola 1993, Wilson *in press*). It is suggested therefore that the chances of preservation of highstand carbonate preservation may be enhanced during areas of highstand siliciclastic progradation.

7.4.4 Carbonate-siliciclastic sequence development in the Fortuna Basin, SE Spain

Sediment input into the Fortuna Basin during the Miocene was controlled by allogenic processes, in particular movement of basin-bounding faults, and uplift and erosion of Betic basement (Montenant 1973, Gracés *et al.* 2001, Krijgsman 2000, Poisson and Lukowski 1996, Santisteban and Taberner 1988). Sediment supply was also strongly influenced by the semi-arid climatic regime, with sedimentation from discrete, high magnitude events (**Sections 5.3.1.1** and **7.2.1.2**). Comparable to the scenario described above for the Vic Basin, lateral shifts in the locus of siliciclastic sedimentation and improved water clarity provided conditions that were locally favourable for carbonate development (**Section 5.3.2**). Smaller, laterally restricted carbonate intervals within the Fortuna Basin did not develop at the same time, indicating an autogenic control on their evolution. Development of massive fringing reef systems (>5 km), exposed on the northern (pers. obs.) and western (Mankiewicz 1995) basin margins would have been strongly influenced by allogenic controls, in particular relative sea-level change.

In the Altorreal area, carbonate-siliciclastic cycles are interpreted to represent the progradation of a high-energy fan-delta into a low-energy, semi-restricted basin with carbonate development on temporarily abandoned delta lobes (**Section 5.3.2**) Carbonate development, in the form of a siliciclastic bioclastic packstone, was initiated on the fan delta top and foresets, forming a small-scale (< 200 m across) shoreline attached ramp-type system (**Section 5.3.2**). In one siliciclastic-carbonate transition, a sedimentary hiatus is indicated from the development of a hardground with a contemporaneous laminar stromatolite unit (**Section 5.3.1.2**). Coral-dominated carbonates that succeed the transitional siliciclastic packstones are relatively 'pure' containing negligible siliciclastic material.

The succession at Altorreal is consistent with the reciprocal sedimentation model described in **Section 7.4.2**. Lithoclastic conglomerates were deposited during the lowstand, with coral development during the highstand. The same interpretation is made for the coeval 'El Desastre' reef-fan delta complex outcropping 10 km to the northwest of Altorreal (Mankiewicz 1995). The lithoclastic conglomerates, interpreted as marine throughout from the presence of oysters and corals, were deposited during a lowstand. This is supported by erosional lower contacts with prodelta marls (**Figure**

5.4a) and coral-dominated carbonates (**Figure 5.16**). Fan delta lobe abandonment and carbonate ramp initiation (**Section 5.3.2**) represents siliciclastic starvation and transgression, although laminar stromatolites and pebble-grade lithoclasts indicate water depths did not increase significantly. The development of coral-dominated clinoforms represents the highstand. Prior to the onset of siliciclastic sedimentation, a period of non-deposition and freshwater inundation, indicated from the presence of columnar stromatolites, contributed to the termination of carbonate development. The erosive nature of the upper contact of the carbonates with the fan-delta conglomerates is consistent with a fall in base level (Holmes and Christie-Blick 1993, Mankiewicz 1995). Changes in water depth however would have been minor (less than 1 m based on the depth of erosion of carbonates) as there is no evidence to suggest movement of the shelf margin break in slope between phases of siliciclastic progradation (**Section 5.3**). Siliciclastics deposited on top of carbonates are coarse-grained sediments deposited within shallow, high-energy conditions.

7.4.5 Sequence development: Calders vs. Altorreal

A summary and comparison of the major controls on sequence development at Calders and Altorreal is presented on **Figure 7.9**. The studied sediments were deposited under different tectonic and climatic regimes, and the resultant successions are different in terms of depositional texture and architecture, and the biota present. However, the mechanisms proposed for the development of carbonate-siliciclastic cycles are essentially the same. Temporary abandonment of a siliciclastic substrate through shifting of the locus of sediment input provided a suitable colonisation site within the photic zone. Initial carbonate production is as a siliciclastic-rich limestone representing the cessation of siliciclastic input and the transformation of conditions towards those more suitable for photoautotrophs. The cessation of carbonate development and switch back to siliciclastic sedimentation caused mass mortality of carbonate producing organisms.

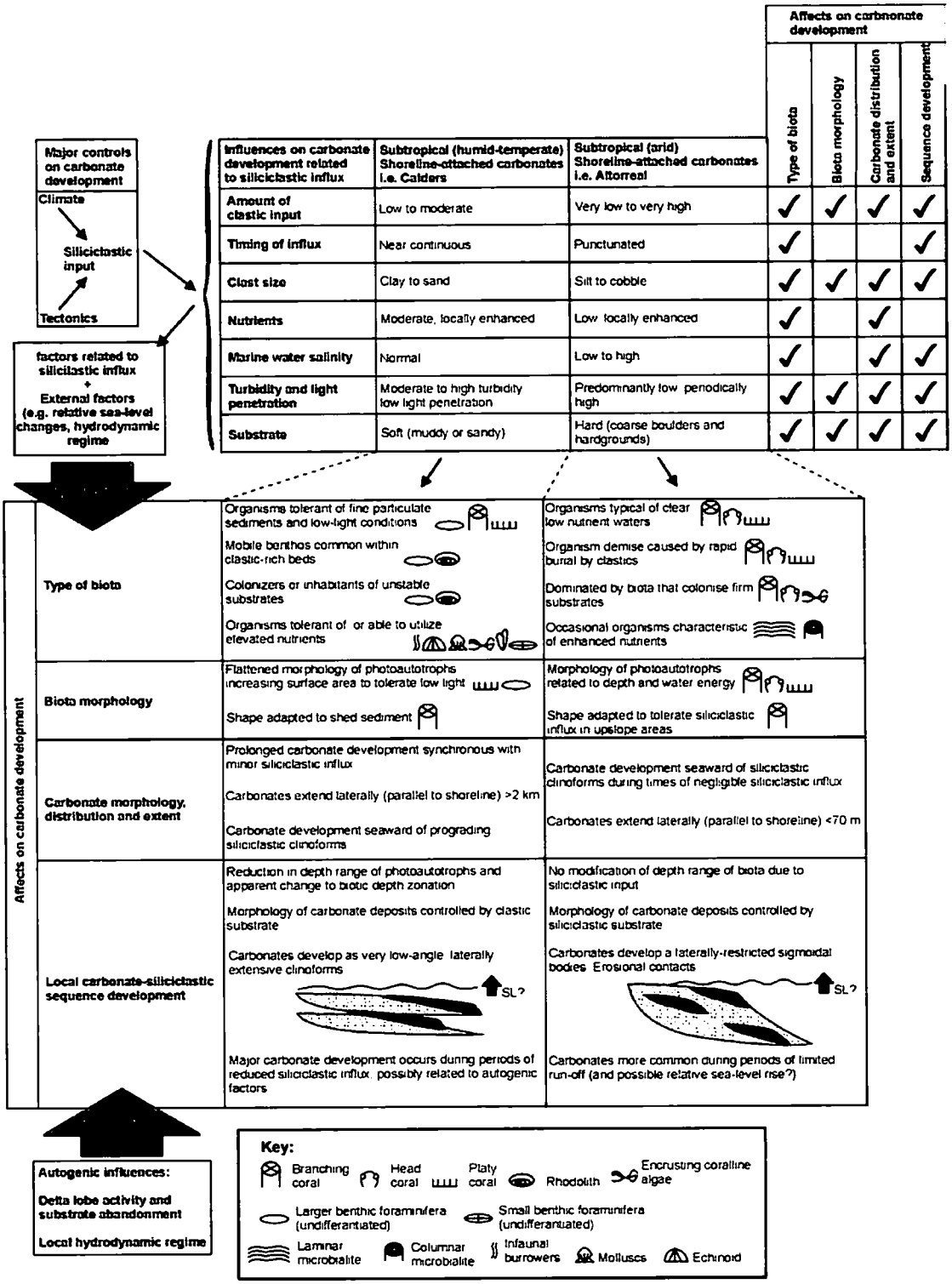


Figure 7.9 Summary of the main controls of carbonate-siliciclastic sequence development at Calders (NE Spain) and Altorreal (SE Spain)

8. Summary

The aim of this thesis was to investigate the development of carbonates within siliciclastic-dominated settings through a detailed analysis of mixed carbonate-silciciclastic successions exposed in NE and SE Spain. A brief review of depositional environments is provided in **Section 8.1**. **Section 8.2** summarises the main effects of siliciclastic input on biota abundances and the influence of siliciclastic input on sequence development is summarised in **Section 8.3**.

8.1 Depositional environments

A summary of facies and depositional environments for the Calders and Altorreal study areas is provided on **Tables 3.2** and **5.1** respectively. A summary of depositional environments identified in the Sant Amanc area is presented on **Table 3.5**.

8.1.1 Calders and Sant Amanc study area, Vic Basin, NE Spain

- ☆ The studied sediments at Calders are Bartonian to Priabonian in age (evident from the presence of *Nummulites bedai*), and comprise part of the Marine Sequence that represents the final marine deposits in the SE Pyrenean Foreland Basin. The studied carbonate siliciclastic succession overlies prodelta marly sandstones of the Vespella Formation (**Section 3.2.1**). The studied section is overlain by fluvial and continental sediments of the Complejos de Calders Formation (**Section 3.2.2**) and the Artès Foramation (**Section 3.2.5**). Partially-restricted marine sediments of the Terminal Complex are also exposed (**Sections 3.2.3** and **3.2.4**).

- ☆ In the Calders study area, 19 facies have been identified (**Section 3.3**). Facies are grouped into carbonate-dominated (**Section 3.3.1**), mixed carbonate-siliciclastic (**Section 3.3.2**) and siliciclastic-dominated groups (**Sections 3.3.3** and **3.3.4**). Siliciclastic-dominated sediments may be fossiliferous or non-fossiliferous. Sediments in the Calders study area occur in 5 facies associations (**Table 3.3**).

- ☆ The sedimentary succession at Calders is divided into 6 siliciclastic-carbonate cycles (**Figure 3.46**). Each cycle represents the colonisation of a siliciclastic substrate during phases of reduced siliciclastic input initially by a foraminifera

and coralline algae-dominated benthic community and then diversification into a coralgal foraminifera-dominated community (**Section 3.4.6**). Two types of siliciclastic to carbonate transition are identified (**Figure 3.61**). The first involves the colonisation of the siliciclastic substrate by *Nummulites* and coralline algae followed by development of coral reef-type facies. The second involves the development of siliciclastic sea grass shoals that supported an array of epiphytic biota, in particular *Gypsina* (**Figure 3.61**). The second transition type occurred in relatively upslope, high-energy areas in comparison to the first transition-type.

☆ It is proposed that sediments exposed in the Sant Amanc area to the immediate northeast of Calders are an as yet unstudied part of the Terminal Complex, as defined by Travé (1992). Four facies are identified (**Section 3.4**). Facies are grouped into coral dominated, mixed carbonate-siliciclastic and siliciclastic facies groups. Interbedded calcareous litharenitic siltstones and sandstones, and porceallaneous foraminifera micritic litharenite facies exposed in the Sant Amanc study area, overlie the upper-most carbonate unit of the Calders succession (**Figure 3.62**) and represent a change from open marine to partially restricted marine deposition.

☆ The siliciclastic to carbonate transition in the Sant Amanc area is marked by the widespread development of larger benthic foraminifera-dominated vegetated shoals (**Section 3.6**). Foraminifera are dominated by *Orbitolites, Rhabdorites* and miliolids with smaller hyaline benthics and molluscs. Oligotrophic conditions persisted in the water column, although locally mesotrophic conditions existed at the sediment-water interface. The baffling of sediment by sea floor vegetation generated minor topography, providing slightly elevated colonisation sites for corals.

☆ Carbonate development in the Vic Basin was terminated in the late Middle Eocene through massive siliciclastic input as the Pyrenean Basin system progressed from its underfilled to overfilled stage.

8.1.2 Altorreal study area, Fortuna Basin, SE Spain

☆ The studied sediments in the Altorreal area are Upper Tortonian in age, and were deposited along the boundary between the Internal and External Zones of

the Betic Cordillera (**Chapter 2**). The studied succession formed prior to the Tortonian Salinity Crisis of the western Mediterranean, and open marine conditions prevailed.

- ☆ In the Altorreal study area, 7 facies are identified (**Section 5.2**). Facies are grouped into siliciclastic-dominated (**Section 5.2.1**), mixed carbonate-siliciclastic (**Section 5.2.2**) and carbonate-dominated (**Section 5.2.3**) groups.
- ☆ Isolated metre-scale carbonate bodies occur within fan delta sediments (**Section 5.2.3**) although carbonate development was not contemporaneous with silciclastic input in the Fortuna Basin. The siliciclastic to carbonate transition is marked by a period of slow deposition and the development of a siliciclastic-rich molluscan packstone. Laminar stromatolites developed in upslope areas (**Section 5.2.3.1**). Coral development occurred along the slope of fan-delta foresets.

8.2 Biota response to siliciclastic input

- ☆ Acid digestion allowed the quantification of the total non-carbonate content of samples (as a weight percentage) and an estimate of the siliciclastic sediment input during phases of carbonate development. Plots of the weight percentage of non-carbonate material against the relative abundance of organisms has allowed the effects of varying amounts and grainsize of siliciclastic input on biota to be made. Sediments from the Calders area (NE Spain) were chosen for acid digestion analysis since these sediments contained *in situ* carbonate-siliciclastic sediments (*sensu* Mount 1984).
- ☆ Variations in siliciclastic input results in changes in the benthic community through the creation of essentially different depositional environments. The punctuated nature of siliciclastic input in the Calders area resulted in the development of two distinctly different environments: prograding siliciclastic shelf and coral reef. Each environment contains a distinct benthic community.
- ☆ The scarcity of biota within siliciclastic horizons is attributed to the rate and amount of siliciclastic input. Colonisation of siliciclastic substrates by immobile calcareous organisms was inhibited largely due to immobile substrates. *Gypsina* was able to survive in a silciclastic-dominated setting

through adapting to an epiphytic mode of life encrusting sea grass that provided a relatively stable substrate. Large benthic foraminifera such as *Nummulites* are able to tolerate moderate sediment input through their ability to extract themselves.

☆ The relative abundance of corals shows a correlation to both non-carbonate sediment abundance and grainsize. Delicate branching corals were present in sediments containing up to 38 wt. % non-carbonate material, although corals are absent where the non-carbonate grainsize >1000μm. Corals are thought to have grown constratally in these settings as sedimentation rates were high and turbidity was moderate to low (or only periodically high). In low-energy settings where turbidity was high, platy coral dominated sediments formed.

☆ Larger benthic foraminifera, such as *Nummulites* and *Discocyclina,* were often the first organisms to colonise siliciclastic substrates. Larger benthic foraminifera were moderately common in sediments containing up to 65 wt. % non-carbonate and thus were able to tolerate prolonged periods of silciclastic input. Larger foraminfera are absent in sediments where the siliciclastic grainsize >4 mm as coarse grains would have abraded and damaged tests. A relative abundance of larger foraminifera within siliciclastic sediments may be attributed to a lack of competitors (or predators) in a stressed environmental setting, an ability to survive burial for a short period of time, an ability to extract themselves following shallow burial and an ability to colonise barren substrates.

☆ Coralline algae are present in sediments where the non-carbonate content is up 38.5 wt. % and are most abundant at concentrations of 10 to 30 wt. % (**Section 6.3.1.2**). In transitional settings (and in association with larger benthic foraminifera), coralline alage occur mainly as rhodoliths, a growth form adapted to immobile substrates. Rhodoliths may provide stable substrate colonisation sites for sessile encrusting organisms in unstable sedimentary settings with an immobile substrate. Irregular rhodoliths develop where large encrusting foraminifera such as *Gypsina, Haddonia* and *Fabiania* have been incorporated.

8.3 Influences on sequence development

- ☆ It is difficult when studying mixed carbonate-siliciclastic systems in the rock record to isolate a simple mechanism for sequence development. Allogenic factors (the climatic and tectonic regimes and sea-level changes) and autogenic factors (hydrodynamic regime, type and diversity of biota, nutrient flux, salinity of marine waters) will influence mixed carbonate-siliciclastic sequence development.

- ☆ The sites and duration of reef development at Calders and Altorreal were influenced to variable extents by the climatic regime, which in turn influenced the rate, magnitude and composition of sediment input. In subtropical semi-arid settings, terrigenous sediment is often locally supplied to the shelf via ephemeral fluvial or wadi systems during infrequent storm events. Consequently, reef development can occur in shoreline attached, shallow water areas such as Altorreal. In relatively humid settings such as Calders where runoff and precipitation are high, the shelf may be characterised by near-continuous siliciclastic and freshwater input. Carbonates tend to develop in relatively offshore positions of the shelf during phases of siliciclastic abandonment, although resuspension of chemically weathered sediments may limit carbonate development by inhibiting photosynthesis.

- ☆ Modern and ancient mixed carbonate-siliciclastic successions accumulated under a range of tectonic regimes including extensional, compressional, passive margin and foldbelt settings. The effects of tectonism on sequence development at Calders and Altorreal are not strongly evident. The Fortuna Basin developed within an extensional setting dominated by block faulting. Uplifted Betic basement provided a nearby source of sediment that was transferred to the shelf through fan delta systems. In other locations in the Fortuna Basin, it has been demonstrated that phases of reef development coincided with tectonic quiescence (Lonergan and Schreiber 1993). The Calders and Sant Amanc successions developed on the southern passive margin of a foreland basin system. Carbonate development on foreland passive margins is common as it is typically a site of minimal siliciclastic input. However, siliciclastic input to the Vic Basin is attributed to the reactivation of Mesozoic fault systems in the hinterland. The Catalan Coastal

Range shed siliciclastic sediments into the southern foreland basin margin through major alluvial fan and fan delta systems.

☆ A seemingly underestimated factor in understanding the relationship between carbonate production and siliciclastic sedimentation is the local hydrodynamic regime. Under situations where the sediment influx is not high magnitude such as flash flood dominated deposition characteristic of the Red Sea coast, it is not so much as how much sediment reaches the coastal system but what happens to it when it gets there. It has been demonstrated that coral reefs tend not develop close to point sources of siliciclastic sediments such as river mouths. The Abrolhos (Brazil) and Paluma Shoals (central GBR) reefs developed only a few kilometres from the shore in areas where fine fluvially-derived sediments are re-suspended by the action of longshore currents. Longshore currents also influence the morphology of the reef.

☆ Autogenic processes on timescales of 10^3 to 10^5 years are thought to have been the most significant influence on the development of carbonate-siliciclastic cycles at Calders and Altorreal. The dominant influence on sequence development is thought to have been delta lobe switching that provided temporary sites shielded from siliciclastic input. Autogenic factors such as the hydrodynamic regime influenced the generation of turbidity. Coral development was influenced by particulate non-carbonate sediment input during phases of siliciclastic abandonment. Turbidity was generated through a combination of re-suspension of locally-derived muddy siliciclastic sediments, and an input of sediments through the action of longshore currents.

☆ In mixed carbonate-siliciclastic successions, the conventional interpretation of depositional sequences is that they are due to reciprocal sedimentation in response to relative sea-level changes. Accordingly, transgressive and highstand deposits are carbonates and lowstands are represented by siliciclastics. Biogenic carbonate production can only occur during a transgression if there is a decrease in siliciclastic input and an improvement in water clarity to compensate for the increase in water depth. Highstand shelves can still be subject to turbid conditions through the action of longshore currents and highstand progradation of siliciclastic shelf systems.

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